EQUATORIAL WEATHER

I. E. M. WATTS

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With particular reference to Southeast Asia

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UNIVERSITY OF LONDON PRESS LTD.
WARWICK SQUARE, LONDON, E.C.4

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Preface

This book aims to furnish an account of the principles of analysing equatorial weather. It seeks to serve needs in colleges, and to provide aviators and seamen with guidance on the weather they must deal with in low latitudes. It is assumed that the reader is acquainted with the elements of weather observing, the standard pressure patterns, and with the types and modes of formation of clouds and thunderstorms.

Many of the principles involved in equatorial weather analysis are new and still in the speculative phase. The scheme of analysis described here is, in a general form, applicable to the whole equatorial region; the application is shown by detailed examples from Equatorial Southeast Asia, where the author has had special experience in the methods of analysis which have been in general use.

Because the conditions of higher latitudes have an extensive literature, the author repeatedly draws parallels and contrasts between them and those of the Equator, so that the readers may see the extent to which meteorological procedures of the temperate zone apply at the Equator.

The presentation is made logically against a background of the principles before attempting practical questions of weather analysis. Because a practical discussion of instances is one of the aims of this book, those theoretical aspects already well-covered in British and American text-books are only briefly expounded. The mathematical theory involved has been reduced as much as possible in the interest of the widest range of readers.

The author is indebted to Prof. E. H. G. Dobby of the University of Malaya for his invaluable guidance and editing throughout the preparation of the book. Dr. C. A. Lea, Director of the Malayan Meteorological Service, has given permission to use much official material. Mr. I. G. John, of the same service, generously gave many helpful suggestions.

Singapore I. E. M. WATTS

Acknowledgments

The author and publisher wish to express their thanks to the Controller of H.M. Stationery Office and the Director of the Meteorological Office, London, for permission to reproduce the data in Fig. 38; to the Royal New Zealand Air Force for permission to reproduce Plates I, III, IV and V; to Captain W. H. B. Wood for permission to reproduce Plate VIII, and to Maureen Verity for drawing the maps and diagrams.

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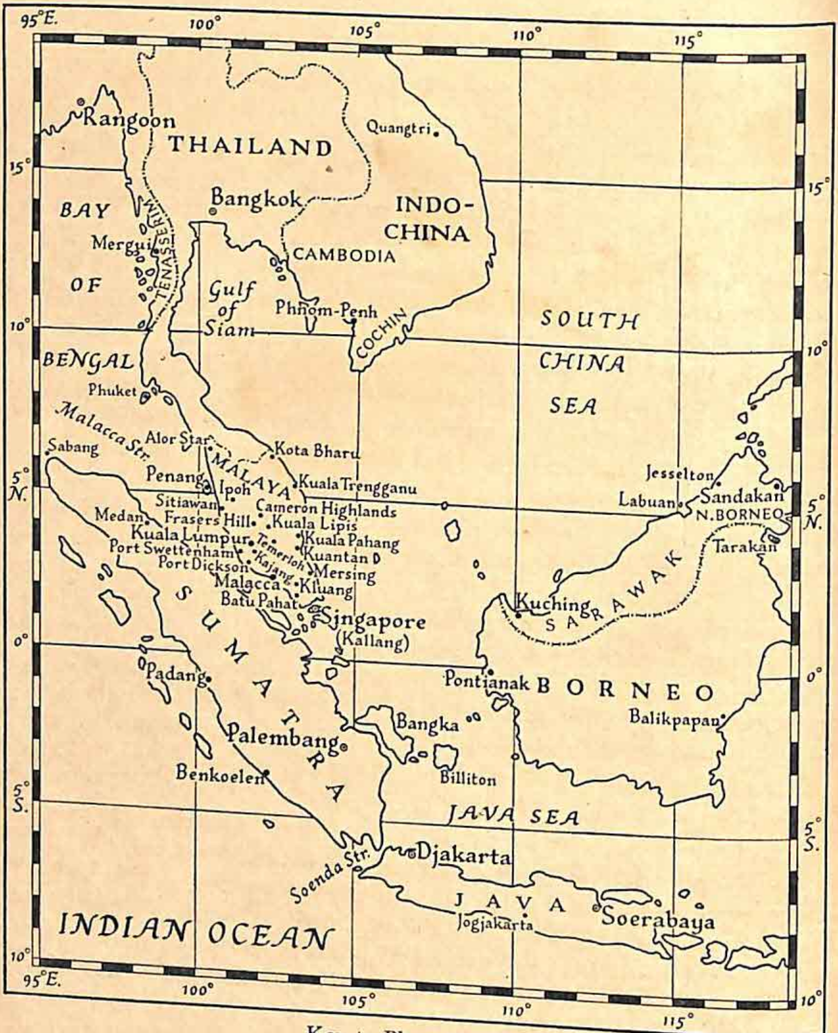


Key to Place Names

CHAPTER I

Equatorial Air Circulation

1. Introduction

When we speak of the equatorial region we refer to one bounded by latitudes 10° North and South. Into other parts of the tropical region, both 'fronts' from temperate latitudes and 'air-stream boundaries' of equatorial origin intrude. Since, however, many currents of air over the Equator originate in much higher latitudes, it is necessary to consider conditions outside the equatorial region proper in studying certain aspects.

The meteorology of low latitudes has been much neglected, so that its study has lagged far behind that of middle latitude weather-forms. At first, disinterest was largely due to the misconception that equatorial weather has no regularity of form—that showers and thunderstorms develop and dissipate indiscriminately near the Equator.

From 1939 to 1945 the needs of war demanded that meteorologists examine conditions at low latitudes, and for a while their interest was greatly stimulated. The accepted theory then was that, over the oceans, most precipitation occurred along a single line, which moved north and south with the seasons and separated the Trade-wind streams of either hemisphere. This line was called the 'Intertropical Front.'

It did not prove very satisfactory as a basis for forecasts because, although rain was sometimes instantaneously localised along it, this line appeared to jump from place to place without regularity. While the conception of an 'Intertropical Front' proved useful on occasions, to trace its movements from one day's chart to the next was difficult.

Later the theory was rejected. Some meteorologists, feeling that too many years had been wasted trying to fit what were fundamentally middle-latitude ideas to the problem of low latitudes, discarded the concept of the Front and abandoned the isobaric weather charts. To replace these, they sought firstly a mathematical expression for the relation between pressure and wind, and secondly a correlation between cloud development and the pattern of the wind-field.

The former did not prove fruitful; the latter led to a method

useful in regions where observations of wind are plentiful. Isobars crept back into low-latitude meteorological practice, and a type of analysis analogous to the 'frontal' method of higher latitudes is still in common use. Both the old and new systems have something to offer. The meteorologist adopts a combination of both, and the practical applications are presented here.

2. The General Circulation

Some features of the winds over the equatorial region have long been known. People of the Southern Pacific knew that there was a steady easterly wind over parts of the tropics, and used it for large-scale migration among oceanic islands. European sailors later took advantage of these easterlies in their voyages to the Americas. Even today Bugis traders sail fleets of small craft down the easterlies from the Celebes to Singapore each July, and await the January westerlies for the voyage home.

The use of steam-power during the nineteenth century encouraged travel in previously unfrequented waters, and there was gradually built up a picture of the prevailing winds over the entire globe. In each hemisphere a belt of light variable winds at mid-latitudes separated a moderate easterly tropical flow from a belt of strong westerlies which extended towards the Poles. The westerlies, found at about latitude forty, were commonly known as the 'Roaring Forties,' and the easterlies were termed the 'Trade Winds.' Near the Equator, between the Trades of the Northern Hemisphere and those of the Southern Hemisphere, there was said to be a region of light winds and calms known as the 'Doldrums.'

As knowledge of the prevailing winds increased, inquiry turned to the physical processes affecting them, and today, even though there is much to be learned, we know what contributes to the general circulation.

The flow of air over the earth is started by the sun's heat. Owing to the earth's curvature and the angle of incidence of the sun's rays, temperatures are highest near the Equator and lowest at the Poles. Air over the heated region expands upwards and outwards, creating a fall of pressure at the lower levels. If the earth did not rotate, the ensuing latitudinal gradient of pressure would result in a low-level flow from the Poles to the Equator, as northerlies in the Northern Hemisphere and southerlies in the Southern Hemisphere. These would rise upon entering the tropics, and at an upper level return polewards, where they would subside to rejoin the surface flow to the Equator.

The actual circulation is far more complex. When air moves over

the earth, the rotation causes it to be deflected to the right in the Northern Hemisphere and to the left in the Southern Hemisphere (see Chapter VII), so that, in the absence of other forces, longitudinal flow from Poles to Equator is not possible. Hence the air flowing away at high levels from the tropics is deflected to cross the temperate latitudes as a southwesterly wind (in the Northern Hemisphere), and the low-level flow towards the Equator is deflected so that it approaches the Equator from the northeast. These Northern Hemisphere flows are mirrored in the Southern Hemisphere, where the surface Trade-winds are southeasterlies and the upper flow northwesterly.

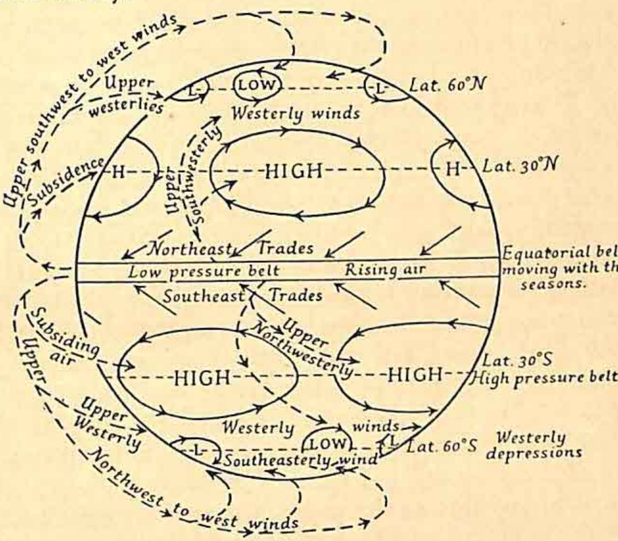


FIG. 1 The General Circulation

Only part of the upper northwesterly and southwesterly flows reaches the Poles. Much air subsides from these upper streams near latitudes 30° N. and S., and the consequent piling-up of air creates two belts of high atmospheric pressure around the earth (Fig. 1). On the sides of the high-pressure belts nearest the Equator are the Northeast and Southeast Trades, while along the poleward edge of each belt is the region of strong westerlies.

At the Poles, low temperatures cause surface pressures to be high. Air subsiding there and outflowing will be deflected, so that around the Arctic Zone is a region of low-level northeasterlies, and around the Antarctic a band of southeasterlies. There remain two regions of lower pressure along latitudes 60° N. and S. respectively, each of which is bounded by the Polar easterlies on one side and the middle-latitude westerlies on the other (Fig. 1).

3. The Pressure Systems

There are complications to this picture. The northeasterlies and southeasterlies around Polar regions do not constitute a steady current so much as an average condition. In the Northern Hemisphere, the northeasterlies occur in thrusts of fast-moving Polar air which, encountering tropical air-masses, stimulate the formation of an intermittent train of 'depressions' in the zone between 40° N. and 60° N. The cold air may pass through the westerlies of the 'forties,' divide the high-pressure belts farther south into separate eastward-moving 'highs,' and at times even enter the tropics.

Sometimes small 'cold anticyclones' form in the Polar outflows, but these are comparatively short-lived and go to swell the larger migratory anticyclones on entering the high-pressure belt. The important cold anticyclones form over the great land-masses in winter. When temperatures fall during the winter months, the air overlying continents increases in density, pressures rise and an anticyclonic flow is set up, as in the 'Siberian High' built up each winter. Weaker anticyclones are formed thus even in tropical countries, and India is occasionally covered by a shallow high-pressure area during December and January.

Highs and Lows at Equatorial Latitudes

Now and then weak high-pressure areas are found near the Equator. Pressure differences from place to place are very small about the equatorial region, and even with isobars at intervals of 1 millibar, the centres of high and low pressure are indistinct. There is no circulation round an equatorial high-pressure area, and, because the deflecting force due to the earth's rotation is small there, the air tends to flow out at right-angles to the isobars. Long-period subsidence is unknown near the Equator; anticyclonic areas disperse rapidly, and fine weather is not necessarily associated with them.

Very shallow lows are also sometimes indicated by the isobars in equatorial regions. In these the winds tend to blow inwards at right-angles to the isobars. Depressions are unusual at equatorial latitudes, but when they do occur there is usually cloud development and precipitation.

Tropical Cyclones and Tropical Storms

Revolving storms are common in parts of the tropics farther from the Equator. A definite circulation moves round their centres of low pressure, with wind strengths from a few miles per hour in shallow depressions to well over 70 m.p.h. when central pressures are extremely low. Differences between tropical depressions, storms

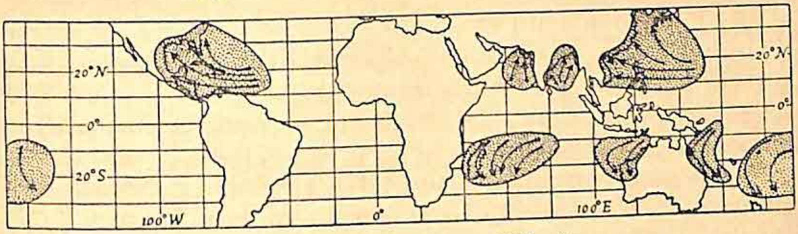


FIG. 2 Cyclone Regions and Tracks

and cyclones are not easily definable but one accepted definition is as follows[1]:

(1) A tropical depression has low pressures within a few closed isobars, and either lacks marked circulation or has winds below 33 knots (Beaufort 7).

(2) A tropical storm is one with several closed isobars and a wind-circulation from 34 to 63 knots (Beaufort 8–11).

(3) A tropical cyclone has a circulation round very low central pressures and winds over 64 knots (Beaufort 12).

These terms are often used loosely and sometimes all circulations around a low are termed cyclones.

Tropical cyclones and storms form only outside the belt 7° N. to 7° S. approximately, and mostly outside 10° N. and 10° S. They mostly form in late summer and over tropical oceans except the South Atlantic, but they never form over continents. The areas commonly affected are shown in Fig. 2.

Of several theories concerning their formation, none has yet proved completely satisfactory. Cyclones certainly require a plentiful supply of moisture, and for this reason their incidence is limited to the regions where the highest sea-surface temperatures are found; that is, to the western regions of the tropical oceans in the late summer. Another favourable condition is when the isobars are wide apart, because then there is weak air-flow and ground heating is at its greatest. Some observers explain that cyclones form on the boundary between two currents of air moving from different directions; others consider that the centre forms at the meeting-place of three distinct streams whose boundaries are quickly lost in the violent rotation about the cyclone.

The first sign of a cyclone is a pressure fall of 2 to 3 millibars in twenty-four hours over a section varying in diameter from two hundred to six hundred miles. Almost at once begins a circulation of freshening winds, clockwise in the Southern Hemisphere and counter-clockwise in the Northern, which blow slightly across the

[1] For this and following references see Bibliography, pages 217–220.

isobars towards the centre at low levels. After a few days, the pressure drops rapidly in the central section and winds may rise to gales. The effective diameter of a mature cyclone is generally from 100 to 300 miles; central pressures approximate to 980 millibars, the lowest record being 887 millibars (reported in a cyclone east of the Philippines in 1927).[2] Winds at low levels may be of devastating force, and velocities of over 100 m.p.h. have been recorded.

Once a fairly vigorous circulation has been set up, a cyclone (in the Northern Hemisphere) moves towards the northwest, with central pressures dropping as it goes. Northerly to northeasterly gales and rapidly falling pressures precede the cyclone. High wisps of cloud gradually thicken and lower, after which heavy rain begins. A marked swell spreads over the sea far ahead of the boundary of the cyclone. The wind suddenly ceases in the 'eye' of the storm, which can be as much as fifty miles in diameter, and the cloud layer often shows breaks. As the centre passes on, gales from the south or southwest set in and pressure starts to rise, while decreasing rain and wind mark the departure of the system. Thunderstorms commonly accompany the cyclone.

Despite the strong winds, the movement of the system is only 10 to 15 m.p.h. at first but increasing. When the centre passes latitude 20° N., the speed of travel increases and the direction usually changes to a more northeasterly path. The point at which the direction of motion of the cyclone changes from being westerly to easterly is known as the 'point of recurvature.' The system then usually begins to fill up; that is, pressure rises in the centre. Tropical cyclones may pass far into temperate latitudes, when their speed of travel may increase to 30 m.p.h.; they are then usually 'decaying' and their wind-speeds diminishing. The recurvature of cyclone tracks in the Southern Hemisphere is in the opposite sense, centres moving first towards the southwest and later towards the southeast.

A cyclone usually fills rapidly once it passes over land, for two reasons: firstly, friction with the ground unbalances the forces acting on the system and an increased volume of air flows across the isobars towards the centre; secondly, continuation of the system depends on an adequate supply of moisture, which is only forthcoming while it is passing over the sea.

4. Cyclonic Regions

Tropical cyclones are given a variety of names, depending on locality. In the Southwest Pacific and the Bay of Bengal the term 'tropical cyclone' is in use; they are known as 'typhoons' in the China Sea, as 'willy-willies' over Western Australia, and as 'hurri-

canes' in the West Indies and in the South Indian Ocean. Few cyclones or storms of any intensity enter Equatorial Southeast Asia, but much of the neighbouring area is affected.

Bay of Bengal

Over the Bay of Bengal about thirteen storms form each year. There is an average of one per month from May onwards, increasing to two per month in August and decreasing again to one each month by December. Few occurrences of storms, depressions or cyclones have been reported in January, February, March or April, and the few which have occurred from January to March have either filled up in the Bay or crossed the Indian coast near Madras.

From April to May the storms move mostly to the northeast and cross the coast between Rangoon and Calcutta; from July to September directions of travel are more to the north or northwest towards the Ganges Delta, with an increase of frequencies on the coast from there to Madras. In November many storms fill while still at sea, and directions of travel vary so greatly that they may cross the coast evenly from Rangoon to southernmost India. The infrequency of reported storms in December is due to the depressions filling up in the Bay, and the occasional storm which does cross the coast may do so either between Rangoon and Calcutta or south of Madras.

China and the Philippines

The number of typhoons occurring in the region from the Marianas across the Philippines to China (from 5° to 30° N. and from 105° to 150° E.) is about twenty per year.[3] Frequencies are negligible from January to April, although isolated occurrences have been recorded. An average of one per month is experienced in May and June. Over the period July to October there are about three to four per month, thereafter decreasing to one per month by December. A great number of lesser storms occur inalso the area.

Typhoon centres normally develop between the Marianas and the Philippines. They mostly move west or west-northwest at first and, after crossing the Philippines, the larger portion recurve and travel northeastward, or northward and often reach Japan. The remainder continue to move on their initial track to the west-northwest, and pass to the coast of China between Hong Kong and Hainan; isolated typhoons have travelled along 10° N. to enter Indochina.

Typhoons cause considerable damage each year to buildings and shipping along the Chinese coast. Hong Kong, between the months

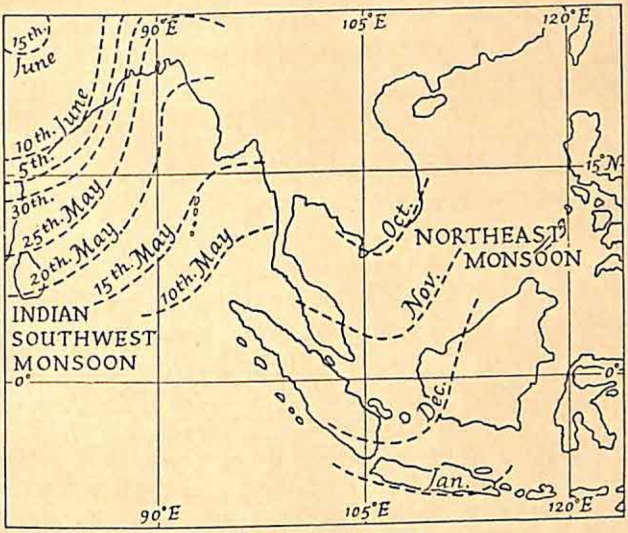


FIG. 3 Date of Onset of the Monsoons

outflow of air from the Indian Ocean is compensated by a continuous inflow from the Southeast Trades of the Southern Hemisphere. The Southwest Monsoon advances across India at much the same time every year, and its arrival marks the beginning of the rains. The mean positions of the monsoons are shown in Fig. 3. It is not possible to give mean positions in other parts of the region as the Southwest Monsoon does not really exist as such south of latitude 5° N., and it is barely distinguishable from other currents over Indochina.

During the northern winter, low temperatures over Siberia cause successive anticyclones to form, and the outflow from these is a broad semi-permanent northeasterly stream over eastern Asia and as far south as Indonesia. This is the Northeast Monsoon. India also has its Northeast Monsoon, fed partly by outflow from the high-pressure north of India and partly by the weak anticyclonic circulation temporarily centred over India itself.

The onset of the Northeast Monsoon initiates the rainy season in all places exposed to the north between China and Java.

The advance is not very regular, so that the mean positions shown in Fig. 3 are less representative than those of the Southwest Monsoon.

6. Wind Systems at Low Levels

The air currents over Southeast Asia and the associated weather vary in four seasons: the Southwest Monsoon, the Northeast Monsoon, and two inter-monsoon seasons of lighter winds.

of July and November, was affected by 6% of the total number of those traced in the whole area over a period of fifty years.[3]

Western Australia

The willy-willies of the West Australia coast originate in the Timor Sea between Timor and Darwin or farther south. They move first to the southwest at 10 m.p.h., but mostly recurve in latitudes 20° to 25° S. when they accelerate and pass inland. Some have reached as far west as Cocos Island before recurving. There are only one or two well-developed willy-willies each year, yet they may cause severe damage. They occur in the period January to March, and occasionally in December and April.

5. Monsoons

The conception of an equatorial low-pressure belt moving north and south with the seasons and with converging Northeast and Southeast Trades on either side of it is a very simplified way of describing conditions at the Equator. Heavy rain or frequent showers and thunderstorms are usually found within the equatorial belt, and for years this bad-weather zone has been called the 'Intertropical Front.' The expression is gradually losing popularity, as it is found that the rain occurs not in a single zone but in two or more separate lines.

In our discussion of the idealised general circulation we have omitted one very important feature—the occurrence of the monsoons. During the summer, the large land-masses become heated. As the overlying air is heated, it expands and pressure falls, and this deficit of pressure is relieved by an inflow from other regions. Similarly, the great cooling of the continents during winter lowers the temperature of the overlying air to much below that of the air overlying the oceans; the increased density leads to higher pressure and a current of air flows seaward. These large-scale seasonal currents from land to sea and from sea to land are known as 'Monsoons.'

Monsoons occur in parts of Africa, in Northern Australia and to a slight extent in portions of North and South America; the most notable examples are found over southern and eastern Asia, where their effect is so great that they completely overshadow the general circulation. Over Equatorial Southeast Asia in particular, the Trade-winds are frequently diverted by or even replaced by the monsoons.

During the period May to September, a broad southwesterly stream of air termed the Southwest Monsoon flows from the South Indian Ocean across all India and sometimes as far as China. The

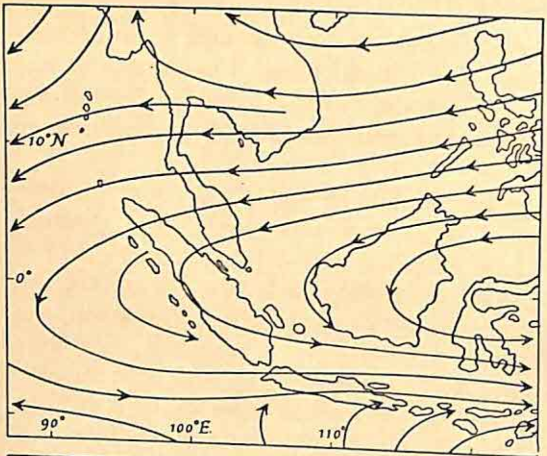


Fig. 4
Surface Winds—January

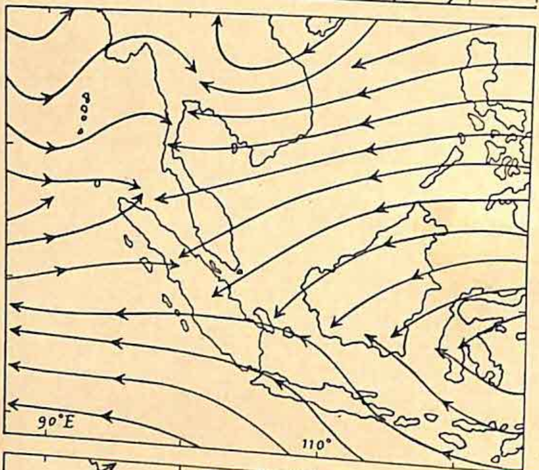


Fig. 5
Surface Winds—April

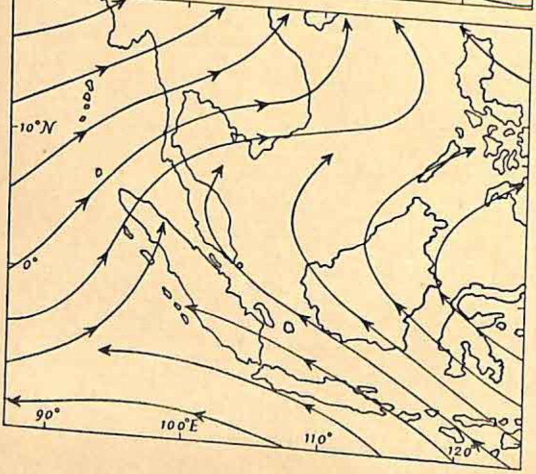


Fig. 6
Surface Winds—July

In January, the Northeast Monsoon has already extended far south and covers the region from Indochina, Thailand and the Philippines to Java (Fig. 4). Over the Northern Bay of Bengal are northerlies from India which, in the southern part of the Bay, merge with the Northeast Monsoon.

South of the Equator the monsoon is frequently diverted to flow in a near-westerly direction. Sometimes this diversion is gradual over the entire stream, while at other times the monsoon air becomes westerly far out in the Indian Ocean. In the latter case the true northeasterlies and the westerlies flow side by side in the Sumatra–Borneo region, and bad weather frequently develops between them. In the south the southeasterlies, which constitute Southwest Pacific Trades, are turned to westerlies or to easterlies as they meet the diverted monsoon. This boundary between the streams is also a region of much cloud and precipitation.

During the months which follow, the Siberian anticyclone slowly loses intensity, and the southernmost boundary of the northeasterlies retreats northward, accompanied by an advance of the Southern Hemisphere Trades. By April, the Siberian outflow is negligible, and any northeasterlies over the South China Sea are then merely an extension of the North Pacific Trades. The Southwest Pacific Trades have reached the Equator (Fig. 5) and westerlies have developed west of Sumatra between the Equator and 5° N. These new westerlies from far out in the Indian Ocean are probably turned Southwest Pacific Trades, and they represent the commencement of the Indian Southwest Monsoon. The other westerlies in the northern Bay of Bengal (Fig. 5) are a weak flow originating in the far northwest.

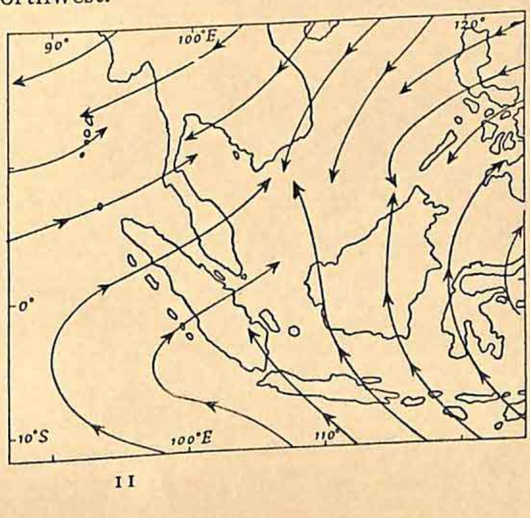


Fig. 7
Surface Winds—October

From May onwards, the Indian Southwest Monsoon rapidly crosses all the Bay of Bengal, Thailand and Indochina. The Trades of the Southwest Pacific also advance northwards in the region east of longitude 100° E., so that by July they have reached latitude 10° N. (Fig. 6). The Trade-winds of the Northern Pacific are kept out of the China Sea by the Southwest Monsoon, but occasionally these Trades appear over the Philippines, where they become southeasterly and conform with the monsoon flow.

In August occurs a slackening of the Indian Southwest Monsoon, and thereafter all wind-streams begin to grow lighter.

By October the Northeast Monsoon is developing in the far north, and as it moves southwards the North Pacific Trades are restricted to the Eastern China Sea (Fig. 7). The Northeast Monsoon then slowly advances until by January it covers practically all Equatorial Southeast Asia. As it advances there is a corresponding weakening of both the Indian Ocean southwesterlies and the Trades of the Southwest Pacific.

These seasonal changes are very different from those which might be expected from the planetary circulation pattern. When they are well-developed, the monsoons dominate the circulation over a wide area. As rainfall over most of Southeast Asia is in some way connected with the monsoons, the time of their onset and retreat concerns nearly all the inhabitants of the region. Planting and harvesting are adjusted to the rainy season. Certain districts suffer severe floods at the same time each year, and during some seasons it becomes impossible to continue engineering projects and sea and air services.

CHAPTER II

Observations

1. Value of the Observational Elements at Low Latitudes

Although the elements contained in equatorial weather observations are much the same as those observed elsewhere (Appendix A), the more important elements of a temperate-latitude report are not necessarily so significant at low latitudes.

Pressure Tendency

Pressure tendency (or change of pressure in the previous three hours) is of considerable value in the temperate zone. Bad weather there is usually associated with the approach of low pressure in a depression or trough, so that, if a continuous record of pressure is maintained, the arrival of adverse conditions may be foreseen from any marked drop of pressure. The diurnal variation only ranges about 1 millibar at middle latitudes, but changes of pressure due to the movement of the pressure systems (highs, lows, troughs, etc.) are quite considerable, and may easily be seen on the barograph despite the diurnal variation.

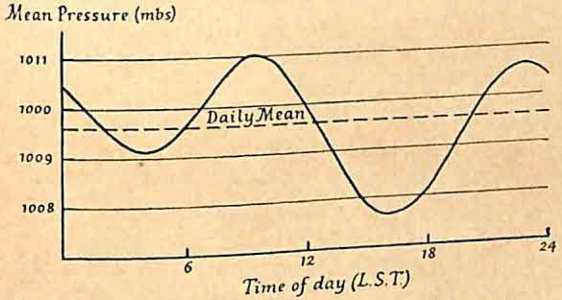


FIG. 8 Diurnal Pressure Variation—Singapore

Near the Equator the daily variation of pressure is much greater —about 3 to 4 millibars. A typical variation may be found in Fig. 8, where the mean pressure at Singapore has been plotted for each hour of the day. The two maxima and two minima are very prominent, and the maximum range is 3·4 millibars. The highs and lows

(excluding cyclones) of the equatorial region are very shallow and pressure differences across them are rarely more than about 1 millibar. Thus the diurnal pressure range masks any changes which might be due to the movement or intensification of the pressure systems, so that pressure tendency has little use as an aid to forecasting. Admittedly, the changes associated with an approaching cyclone are large enough to show up through the diurnal variation, yet because a cyclone covers only a small horizontal area, the drop of pressure at its approach occurs only a short time before the onset of the high winds.

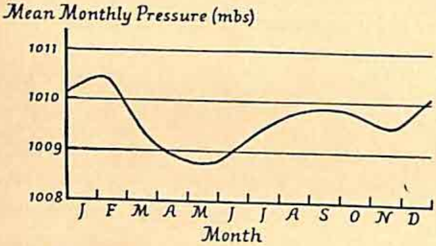


FIG. 9 Seasonal Pressure Variation —Singapore

There are other cyclic changes of pressure besides the diurnal variation. Over Southeast Asia the general level of pressure slowly rises during each of the monsoons and falls to a minimum between them (Fig. 9). The Siberian High slowly builds up from November onwards, and the outflow from it causes a rise of pressure in the equatorial region. There is an overall decrease during March and April as the High loses intensity, but pressures rise again with the onset of the Indian Southwest Monsoon in May.

Short-period surges of pressure also occur. Over Equatorial Southeast Asia rises and falls of pressure regularly take place over a period of five to six days. The large geographic extent of these fluctuations may be seen from the close correlation between the average daily pressures at Singapore, Penang (Malaya) and Kuch-

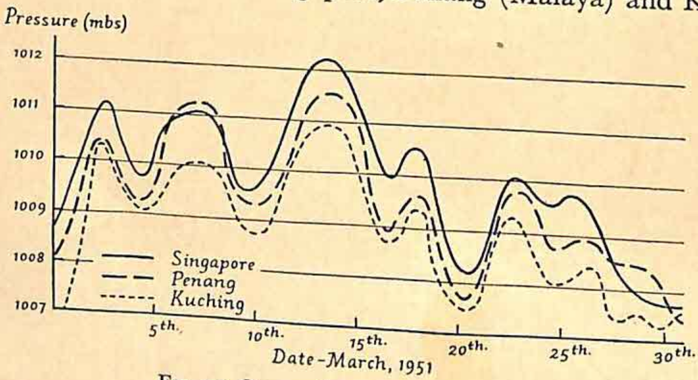


FIG. 10 Comparison of Daily Pressures

ing (Sarawak) during March 1951 (Fig. 10). These three stations, at considerable distances from one another, have practically identical pressure rises and falls from day to day.

Similar large-scale variations of pressure over a few days in higher latitudes are associated with the passage of the migratory anticyclones, in which changing pressure correlates with rainfall. There is no such correlation between rainfall and pressure changes at the Equator. Probably a relationship exists between the short-period variations of pressure at low latitudes and those in the mid-latitude Southern Hemisphere high-pressure belt, because the period of the migratory anticyclones is similar.[4]

Wind

Accurate estimates of wind direction are important to the equatorial meteorologist, whose weather analysis depends upon identifying the general flow of air by comparing wind directions at various places. The best comparison is afforded by using winds at some upper level: whenever upper wind reports are not available the analyst is forced to discriminate between the directions at the earth's surface. Near the great land-masses this is difficult because the winds are much influenced by topography. Reports from small islands and ships rarely suffer from local influences, and are usually more representative of the main flow in their vicinity.

Wind speeds are not of such great significance in the equatorial region, where surface winds are mainly less than 10 m.p.h. except under certain seasonal and local influences and in the neighbourhood of cyclones. Table 1 analyses wind speeds at Kota Bharu (East Malaya).

TABLE 1

WIND SPEEDS AT KOTA BHARU AND THEIR AVERAGE DURATION AS PERCENTAGES FOR EACH MONTH

Wind Speed (m.p.h.)	*Month*												*Year*
	J.	*F.*	*M.*	*A.*	*M.*	*J.*	*J.*	*A.*	*S.*	*O.*	*N.*	*D.*	
Calm	12	11	16	12	11	13	16	17	19	23	16	14	15
1–3	34	38	39	47	43	50	54	50	50	49	51	47	46
4–7	24	25	26	28	36	30	24	25	27	23	25	28	27
8–12	20	22	18	12	10	7	6	7	4	5	8	11	11
13–18	9	4	1	1	0	0	0	1	0	0	0	0	1
19–24	1	0	0	0	0	0	0	0	0	0	0	0	0

The table shows that during 61% of the year the average wind speed is less than 4 m.p.h., and that only during a very small portion of the year are winds stronger than 12 m.p.h. Kota Bharu is exposed to the Northeast Monsoon, and the higher ranges of speed are confined to January, when the monsoon is fully developed over this area.

Temperature

An outstanding feature of the equatorial region is that there is no sharp division into a warm and a cool season. Between latitudes 10° N. and 10° S. the noon elevation of the sun is always high, and therefore the seasonal range of temperature is very small, contributing to the monotony of life near the Equator. The mean monthly temperatures* for Singapore, Kuala Lumpur and Cameron Highlands (a hill-station in Malaya) (Fig. 11) show that the range of the mean monthly temperatures at each of these places is less than 3° F.

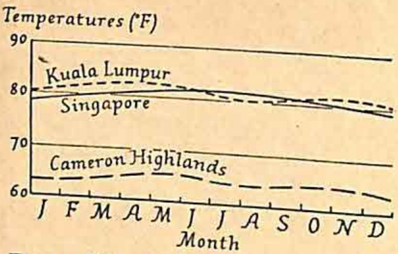


FIG. 11 Mean Monthly Temperatures

The diurnal range† of temperature near the Equator is smaller than that of higher latitudes, and the small fall of temperature at night in low latitudes has enervating effects. The mean daily range at Singapore is about 10° F., and it varies little throughout the year (Fig. 12). Inland stations exhibit a greater diurnal range than coastal stations (Table 2). At Kuala Lumpur, the average diurnal range over the whole year is 18·5° F.—nearly twice that of Singapore. The greater range at Kuala Lumpur has little beneficial effect on human beings because it is brought about by the fact that although Kuala Lumpur maxima are much higher than those of Singapore, the minima of both places are nearly the same. Furthermore, the absolute maxima (highest temperatures ever recorded) for each month are of a higher order at Kuala Lumpur, while the absolute minima are much the same.

At the higher altitudes of Cameron Highlands (4750 ft. above sea-level), maxima are about 14° F. below those of Singapore, and

* The mean monthly temperatures are computed thus: The average of the highest temperatures recorded each day of the month is added to the average of the lowest temperatures recorded each day, and this sum is divided by two.

† The difference between the average of all the early morning minima and the average of all the afternoon maxima over a long period.

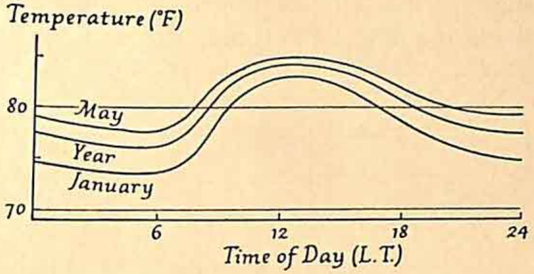


FIG. 12 Diurnal Variation of Temperature—Singapore

minima are 18° below (Table 2). The maximum temperature at Cameron Highlands never reaches 80° F., yet minima only a few degrees above freezing-point have been recorded.

Although the diurnal variation of equatorial temperature and the corresponding variation of dew-point are so small, these elements are not useful in distinguishing between currents of air from different directions at low latitudes, because the currents have such similar temperatures and moisture contents there.

Weather

The recommended[6] spacing for land reporting stations is from 60 to 100 miles apart with a maximum spacing of 300 miles, and most equatorial and tropical countries have developed networks conforming to this standard, though there is considerable doubt whether it is adequate near the Equator. The doubt is illustrated by the facts of Table 3, showing simultaneous reports from four stations on Singapore, an island only 13 by 24 miles. At no time during this day were all four reports identical, despite the nearness of the stations. It is obvious, then, that the customary spacing of 60 to 100 miles must give a poor representation of true conditions, and even stations a quarter of these distances apart might not adequately fulfil the requirements.

The inadequacy arises because much of equatorial weather is local rather than regional. A single cumulonimbus cloud may produce heavy rain over an area varying from less than 1 square mile to 24 square miles. Furthermore, the general flow of air is very light during a great part of the year, causing surface winds to be greatly affected by local influences. In particular, winds near a thunderstorm may change direction and speed frequently within a short time.

There is also doubt whether the existing meteorological vocabulary is suitable for the equatorial region. For instance, in report-

TABLE 2

TEMPERATURE EXTREMES AND SUNSHINE HOURS

Month	*J.*	*F.*	*M.*	*A.*	*M.*	*J.*	*J.*	*A.*	*S.*	*O.*	*N.*	*D.*	*Year*
Singapore													
Mean Max. Temp. (° F.)	85·4	87·6	87·2	87·4	87·2	86·8	86·9	86·8	86·4	86·0	85·9	85·4	86·6
Mean Min. Temp. (° F.)	73·1	73·2	74·2	75·3	76·0	76·3	76·9	76·5	75·5	75·1	74·3	73·7	75·0
Absolute Max. Temp. (° F.)	92	94	94	94	94	93	93	93	92	92	92	91	94
Absolute Min. Temp. (° F.)	67	69	68	70	71	70	70	70	69	70	70	69	67
Sunshine, Hours per Day	5·23	7·02	6·00	5·69	6·21	6·28	6·83	6·52	5·61	5·06	4·78	4·64	5·82
Kuala Lumpur													
Mean Max. Temp. (° F.)	89·3	92·1	92·1	91·6	91·2	90·9	90·4	90·6	90·1	89·5	89·3	88·9	90·5
Mean Min. Temp. (° F.)	71·3	71·5	72·4	73·0	73·1	72·3	71·6	71·6	71·7	71·9	72·2	71·7	72·0
Absolute Max. Temp. (° F.)	95	97	97	96	96	96	98	98	96	96	95	94	98
Absolute Min. Temp. (° F.)	64	66	68	69	69	68	67	68	67	69	69	67	64
Sunshine, Hours per Day	5·83	7·30	6·32	6·09	6·27	6·53	6·31	6·20	5·43	4·90	4·78	5·15	5·93
Cameron Highlands													
Mean Max. Temp. (° F.)	70·5	72·8	73·2	73·9	73·7	73·2	72·7	72·1	71·8	71·3	71·1	70·6	72·2
Mean Min. Temp. (° F.)	56·5	54·0	55·6	57·3	57·8	56·3	55·4	56·1	56·9	57·8	57·6	56·4	56·5
Absolute Max. Temp. (° F.)	77	78	79	79	78	78	77	78	78	77	76	76	79
Absolute Min. Temp. (° F.)	36	40	44	47	48	47	45	48	46	49	44	45	36
Sunshine, Hours per Day	4·32	5·90	4·92	4·63	4·91	5·42	5·12	4·80	4·06	3·61	3·56	3·61	4·57

TABLE 3

CONDITIONS AT SINGAPORE ON 25TH APRIL 1951

Local Time	0730	0830	0930	1030	1130	1230	1330	1430	1530	1630
Kallang .	Thunder Calm	— Calm	— Calm	— Calm	Shower W.S.W., 15 m.p.h.	Rain S.S.E., 10 m.p.h.	Thunder S.S.E., 10 m.p.h.	Drizzle S.E., 5 m.p.h.	— S.S.E., 5 m.p.h.	Rain S., 10 m.p.h.
Tengah .	Rain Calm	— Calm	— W.S.W., 2 m.p.h.	— W., 10 m.p.h.	Shower W.S.W., 15 m.p.h.	Shower N.W., 10 m.p.h.	Thunder W.S.W., 5 m.p.h.	— S.W., 2 m.p.h.	— S.S.W., 15 m.p.h.	— S.S.W., 15 m.p.h.
Seletar .	Thunder Calm	— Calm	— N., 5 m.p.h.	— S.W., 5 m.p.h.	— W.S.W., 5 m.p.h.	Shower S.W., 10 m.p.h.	Shower S., 10 m.p.h.	Shower E.N.E., 10 m.p.h.	Shower E.N.E., 5 m.p.h.	Shower S., 10 m.p.h.
Changi .	— N.W., 2 m.p.h.	— N.W., 2 m.p.h.	— Calm	— Calm	Rain S.S.E., 2 m.p.h.	Shower S.W., 10 m.p.h.	Thunder S., 5 m.p.h.	Thunder W.N.W., 2 m.p.h.	Shower S.E., 2 m.p.h.	Thunder S., 10 m.p.h.

ing 'showers,' the classification of Appendix A provides the terms 'slight,' 'moderate or heavy' and 'violent.' Precipitation causing floods in the equatorial region sometimes falls from solitary cumulonimbus clouds from half a mile to perhaps 5 miles broad. According to definition, such precipitation should be termed a 'shower,' yet even 'violent shower' is quite inadequate to describe the actual precipitation, which may be at the rate of 2 to 3 inches an hour. Falls of this nature might only be described sufficiently by a new term such as 'violent rainstorm.'

2. Surface Synoptic Charts*

A similar deficiency is found in the vocabulary for describing the features introduced in the analysis of the weather charts and upper-wind charts. The technique developed at higher latitudes is known as 'frontal analysis,' and the vocabulary is derived therefrom. Few terms in this vocabulary apply near the Equator. In frontal analysis different currents are distinguished by their temperatures and moisture contents, the currents themselves being termed 'air masses' and the boundary between two different currents being called a 'front.' The air currents of low latitudes do not exhibit measurable differences of temperature and moisture content, for which reason in this volume we will term the currents 'air streams,' not 'air masses.' Furthermore, the behaviour of two equatorial air streams at a common boundary is different from that of two middle-latitude air masses at their boundary, so that the term 'air-stream boundary' will be used in preference to the term 'front' (see Chapter VIII).

Drawing isobars and marking out the areas of bad weather is by no means a simple task near the Equator, where the pressure patterns are weak and isobars widely spaced even though drawn at 1-millibar intervals. The winds are less useful as guidance for drawing equatorial isobars than they prove to be in higher latitudes. This is because equatorial winds tend to blow across the isobars, and the precise mathematical relation between the wind and the isobars has not yet been determined (see Chapter VII). Great care must be exercised in delineating the areas of bad weather, because there is doubt of the representativeness of each observation (the symbols used are shown in Appendix B). For these reasons, the surface chart is less important near the Equator than in middle latitudes.

* Various projections are used for the synoptic weather-charts of middle and high latitudes; the one most suitable[7] for the equatorial region is Mercator's with true scale at latitude 22½°.

3. Upper-wind Charts

The stations observing upper winds in Southeast Asia are spaced at distances from 150 to 400 miles, and the network is supplemented by reports from aircraft. At each reporting hour, separate charts are plotted for various levels, and a useful overall picture of the winds aloft can be derived from maps at 3000, 5000, 10,000, 15,000 and 20,000 ft. Few pilot-balloons are observed to heights greater than 20,000 ft., so that charts for very high altitudes normally contain too few observations to be of much value—a difficulty which may be overcome when more balloon flights are followed by radar.

The plotting is simple. An arrow is drawn from each station to show the direction from which the wind is blowing, and feathers on the arrow denote wind speed—a long feather for each 10 knots and a short feather for each 5. The charts are analysed by drawing stream-lines which show the general flow (at each of the fixed upper levels) from a consideration of the separate reports of wind speed and direction. When streams are discontinuous, the location of the discontinuity is marked. To derive maximum benefit from the upper-wind charts, the stream-lines should be spaced in inverse proportion to the wind speed, but this is not always carried out in practice because winds near the Equator vary so little in speed. The primary use of the stream-lines is to determine the air-stream boundaries at various levels, a process discussed in later chapters.

Upper-wind observations have another use. From their speed and direction may be estimated the degree of horizontal convergence over any area. In this sense, convergence is said to occur when more air is flowing horizontally into some area than is leaving it. Such an inflow must be relieved by up-currents. Because cloud and rain form in ascending air, the areas of convergence are those of present or future bad weather. As tropical rains are essentially local rather than general, it is necessary to make separate calculations of the degree of convergence for many small sections of the whole region, each section being not more than about 50 miles square. The existing upper-wind network is too widely spaced to fulfil requirements.

4. Upper-air Observations

Upper-air observations are made by 'radio-sondes'—instruments which transmit signals of pressure, temperature and humidity at fixed intervals of time as they are carried upwards by balloons. Such reports may be used in two ways.

The observations of an ascent may be plotted on a special chart, the most popular being the 'tephigram' (see Chapter III). It may

be used to determine whether the structure of the atmosphere at any particular time is favourable for up-currents, and therefore for the development of rain. Because the vertical structure of the equatorial atmosphere nearly always favours shower development and as the tephigram will not denote the most probable location of rainfall, the charting has no great value. In middle latitudes, comparing two tephigrams containing ascents plotted for different times or for different places may help to identify two separate currents of air by dissimilarity in their vertical structure. This also is impracticable near the Equator where currents from totally different directions have similar properties of temperature and humidity aloft.

The second use of radio-sonde reports is in the plotting of upper-air charts at certain levels of pressure. Separate charts are usually employed for the 1000-, 750- and 500-millibar pressure levels, and on these charts are plotted (among other elements) the temperature and the true height of the level at each point of radio-sonde observation. From these, contour lines are drawn joining places of equal height, together with the isotherms. The lines of equal height pick out the troughs and lows of the upper atmosphere, while the relation of the height lines to the isotherms shows the speed of travel of the troughs or lows. These upper-air charts are in everyday use in temperate countries. They may also be useful at low latitudes, but over no part of the equatorial region are there sufficient upper-air soundings to enable the method to be tested adequately.

5. Summary

The most valuable observation in equatorial meteorology is that of wind. Small variations of direction are important at ground level, while both direction and speed are significant at upper levels.

CHAPTER III

Conditions in the Upper Air

1. Lapse-rates

Temperature lapse-rates in the troposphere differ from place to place because the air is in constant motion involving currents flowing over warmer or colder regions. A polar stream, thousands of square miles in extent, will have its lower layer heated by the warmer ground; stirring induced by relief will carry the warmth up a few thousand feet, but at higher levels temperatures may remain unchanged. Thus the decline of temperature with height becomes great in the lower strata, and a steep lapse-rate occurs. On the other hand, a tropical stream travelling to more temperate regions will suffer cooling from the land or sea below, until surface temperature in it is nearly as low as that at 4000 or even 8000 ft. —then the lapse-rate becomes very slight. When layers of air of different history lie one above the other, inversions are found and temperature *increases* with height.

If a parcel of unsaturated air is forced upwards it cools at the dry adiabatic lapse-rate of 3° C. (5·4° F.) per thousand feet, irrespective of the lapse-rate in the surrounding air. Upon reaching saturation-point the rate of cooling becomes the saturated adiabatic lapse-rate—about 1·5° C. (2·7° F.) per thousand feet in the lowest layers, but greater aloft.

2. Stability

Examining the lapse-rates helps to determine which air is conducive to up-currents and which will retard them, and this is important because the formation of cumuliform clouds depends on up-currents.

Suppose that observations have been made at each thousand feet. The temperatures, shown in the centre column of Fig. 13*a*, have a lapse-rate of 4·9° C. in the first layer and 4·8° C. in the second—a steep lapse-rate compared with the usual one of 1·7° C. If a parcel of air of temperature 15° C. began to move upwards, it would cool at the dry adiabatic lapse-rate (assuming that the air is not saturated), and at 1000 ft. its temperature would have dropped to 12° C. The rising parcel would now be warmer than the sur-

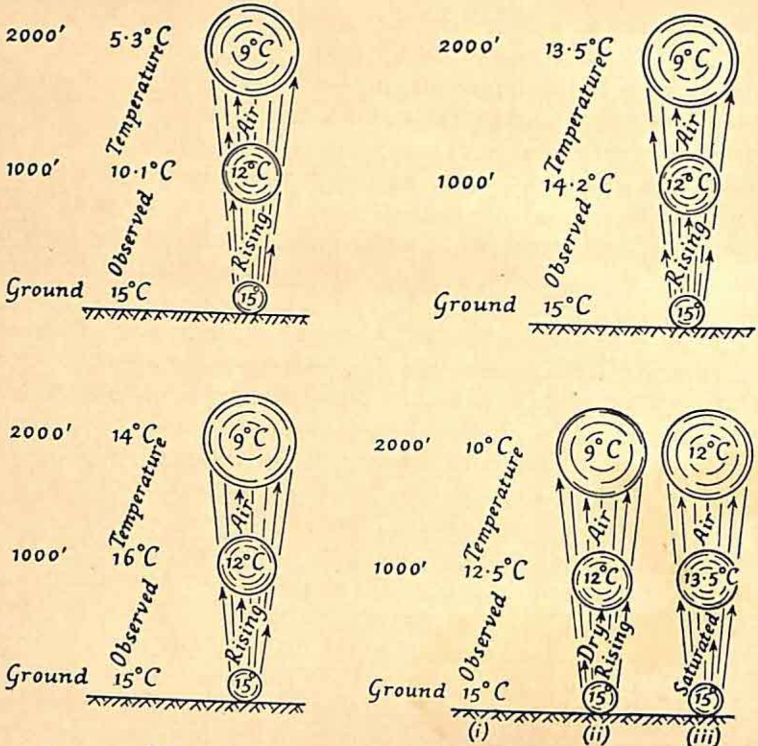


Fig. 13 Stability. The Condition of Rising Air

(*a*) Unstable
(*b*) Stable
(*c*) Inversion
(*d*) Conditionally unstable

rounding air and therefore *continue* upwards. A similar condition is evident at 2000 ft., where the atmospheric temperature is 5·3° C. and that of the rising parcel would be 9° C. Thus, whenever the lapse-rate is steep, upward motions in the air are naturally reinforced, producing an atmosphere called 'unstable' and characterised by convection currents.

Consider now the temperatures in the centre column of the case in Fig. 13*b*. The lapse-rate, 0·8° to 0·7° C. per thousand feet, is slight. Air of initial temperature 15° C., rising to 1000 ft. through this unsaturated atmosphere and cooling at the dry adiabatic rate, becomes colder (12° C.) and denser than the surrounding air (14·2° C.). It ceases to rise and tends to descend to its original position. If by any means it were forced up to 2000 ft., the rising air would become 4·5° colder than the surroundings and further retarded. If the air were saturated, the retarding influence would still

be present but to a lesser degree; in rising to 1000 ft., it would attain 13·5° C., and be still colder than its surroundings. Thus, an atmosphere of slight lapse-rate presents a retarding influence to upward movement and marks a stable condition.

The case of air rising to an inversion requires consideration. The temperatures in Fig. 13*c* have an inversion in the lowest thousand feet, although temperature falls off above it. Whether the air is dry or moist, a rising parcel must become considerably cooler than the warm air of the inversion and upward motion must cease. Hence inversions hinder up-currents.

A special case arises when the observed lapse-rate per thousand feet is intermediate between the dry and saturated adiabatic rates (between 3° and 1·5° C.). An example in Fig 13*d* has an atmosphere with a lapse-rate of 2·5° C. (column i). If this air were non-saturated, any rising portion would cool to 12° C. (column ii) at 1000 ft., and be 0·5° colder than its surroundings. Consequently, its up-currents would be retarded and the air stable. If the air were saturated, however, cooling at the saturated adiabatic rate would result in a temperature of 13·5° C. at 1000 ft. (column iii), reinforcing the upward thrust so that the air would be unstable. Thus, when a lapse-rate is intermediate to the dry and saturated adiabatic rates, the degree of stability depends on the amount of moisture in the air, a dry air being stable and a saturated air being unstable. Such an atmosphere is said to be 'conditionally unstable'—the state most commonly found in the equatorial atmosphere.

3. The Tephigram

Several graphical methods are used for displaying upper-air temperature soundings. The one in general use is the Tephigram, or T-ϕ-gram, where T is the temperature and ϕ is the 'entropy' of the air. When a particle of air warms adiabatically by descent, there is no transference of heat between it and its environment. In direct ratio to the gain of temperature, there is a loss to the store of heat or energy of the particle. The ratio of the two quantities (Energy/Absolute Temperature) remains constant for a sample of air during adiabatic changes, and is termed its 'entropy', being expressed in units of energy per degree of temperature.

In the original tephigram (Fig. 14), the OY-axis has a scale for entropy and the OX-axis a scale for temperature. The potential temperature (θ) is what a specimen of air would attain if brought dry-adiabatically to a standard pressure of 1000 millibars, and is related to entropy as follows:

$\phi = a \log \theta + b$, where a and b are constants. The OY-axis thus may

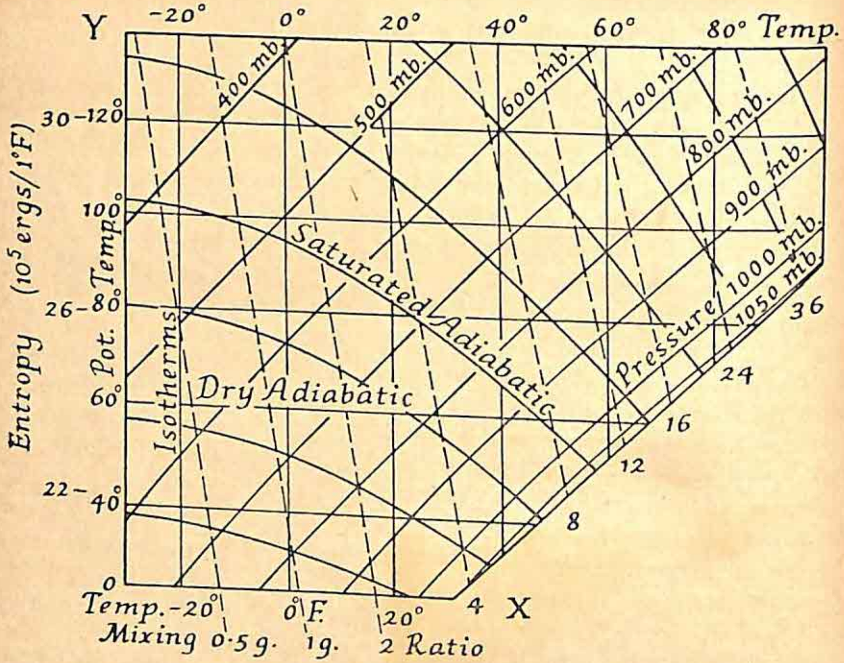


FIG. 14 The Basic Tephigram

also be graduated according to potential temperature, and lines parallel to the OX-axis are dry adiabatic. Saturated adiabatic scales are also drawn on the chart with the assumption that all moisture falls on condensation.

Lines of equal pressure may be shown on the tephigram because pressure and temperature in adiabatic changes are related. $\frac{T}{p^{0 \cdot 288}}$ remains constant for varying values of T and p in a sample of air. When $p = 1000$ millibars, T becomes the potential temperature, and $\frac{T}{p^{0 \cdot 288}} = \frac{\theta}{1000^{0 \cdot 288}}$, from which it follows that $\theta = T\left(\frac{1000}{p}\right)^{0 \cdot 288}$, a formula which enables values of pressure to be determined for different values of temperature and potential temperature, and hence constant-pressure lines may be constructed.

The humidity mixing ratio, which is also shown on the chart, is a measure of the moisture content of the air, and is expressed by $622 \frac{e}{p-e}$ gms. per kgm. of dry air, where e is the partial pressure

of the water vapour contained in the air and p is the total pressure of the air.

The ordinates in the tephigram (Fig. 14) are:

(1) Isotherms—vertical straight lines of constant temperature.
(2) Dry-adiabatic lines—horizontal and straight.
(3) Saturated adiabatic lines—curves sloping upwards to the left.
(4) Lines of equal pressure—slightly curved and sloping up towards the right.
(5) Humidity mixing-ratio lines—the broken straight lines sloping upwards a little to the left of the vertical.

In the improved form of tephigram (Fig. 15) now in more general use, the slightly curved constant-pressure lines are nearly horizontal, which facilitates interpretation of the diagram by those accustomed to temperature-height diagrams. In the modified form, the ordinates are as follows:

(1) The nearly horizontal lines are of constant pressure.
(2) The straight equidistant lines diagonally upwards to the right are isotherms.
(3) The dry adiabatic lines are straight and run diagonally upwards to the left.
(4) The saturated adiabatic lines diverge upwards to the left.
(5) The broken straight lines running upwards to the right represent humidity mixing-ratio.

Plotting on the Tephigram

Four typical plots of upper-air soundings are shown in Fig. 15. Curve ABCD is plotted from the following pressure and temperature data:

Point	*Height*	*Pressure*	*Temperature*	*Relative Humidity*
	ft.	*mbs.*	*° C.*	%
D	3500	905	0·1	50
C	2500	932	4·6	58
B	1500	987	10·4	60
A	ground	1024	15·0	70

The humidity is plotted as a separate curve (KLMN of Fig. 15) as follows. The mixing-ratio line through A shows that the maximum possible amount of water vapour in the air at that temperature

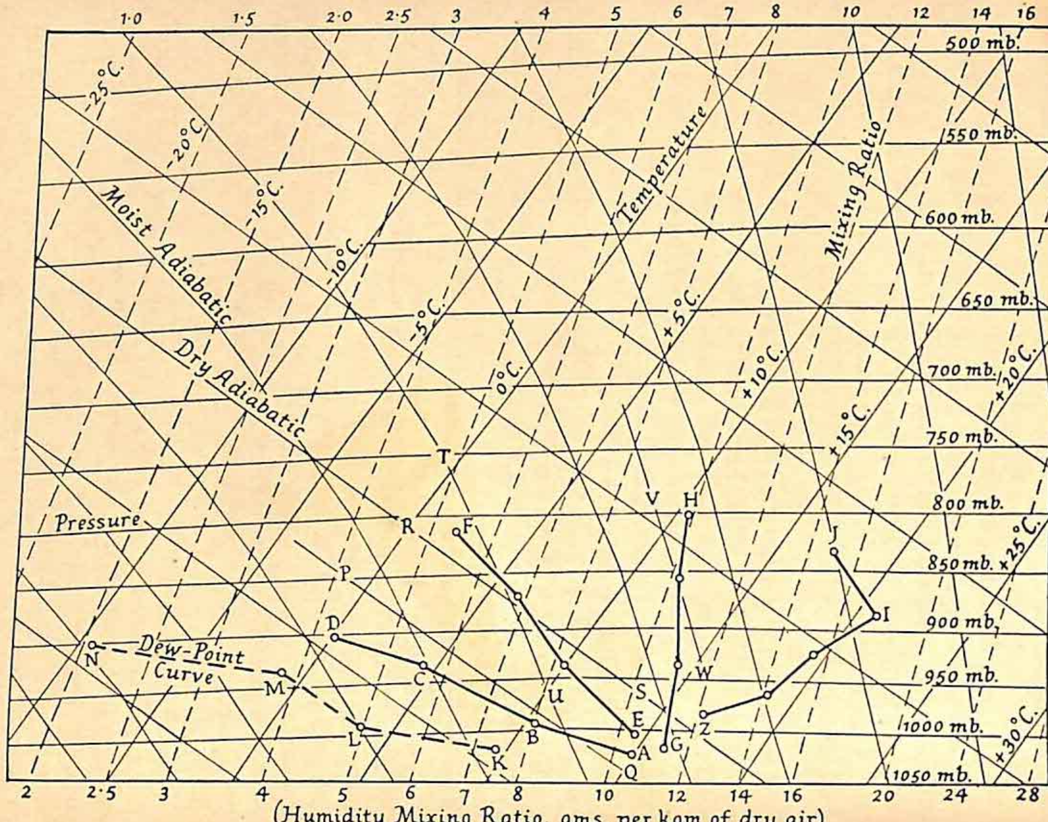


FIG. 15 Types of Lapse-rate

(15°) is 10·6 gms. per kgm. of dry air. As the relative humidity at A is 70%, the actual amount of water vapour is 7·4 gms. (70% of 10·6), and point K is plotted for a mixing-ratio of 7·4 gms. per kgm. at the surface pressure level of 1024 millibars. Points L, M and N are found similarly, and the resulting curve, KLMN, is the 'Dew-point Curve', corresponding to the pressure-temperature curve ABCD.

Interpretation of the Tephigram

Comparing curve ABCD (Fig. 15) with the nearest dry adiabatic line, PQ, it will be seen that the atmosphere under consideration has a steeper lapse than this adiabatic, and hence is unstable. Atmosphere represented by EF has a lapse-rate less than the dry adiabatic RS but steeper than the moist TU, and hence has conditional instability. Curve GH is stable, as it is of less lapse-rate than the saturated VW, while ZI is an inversion.

Use of the Tephigram

The tephigram has two important uses in middle latitudes. In the first place, it displays the distribution of temperature and moisture aloft, so that different air masses may be distinguished by comparing their lapse-rates. This is one of the basic processes in air-mass analysis, and its use is made possible by the large range of lapse-rates occurring in middle latitudes. For instance, the lapse-rate is decreased by low-level cooling in all air currents moving from the tropics, but it is increased by warming in air from the polar regions.

Very different conditions exist at low latitudes, where air currents are slow moving and are greatly influenced by the regions over which they flow. In Equatorial Southeast Asia there are no major land masses, and the most important modifying influence is that of the seas. As isotherms at sea-level are widely spaced in this region, temperatures at low levels become similar in different currents of air. Since most currents come from colder regions, lapse-rates are increased and the resulting convection carries a high moisture content to upper levels. By this process all lapse-rates of temperature and moisture tend to become similar, so that the separate streams cannot be distinguished by use of the tephigram.

Compare the conditions at Seletar on the 5th and 6th of November 1951, for example (Figs. 16*a* and *b*). The upper winds (plotted at the appropriate levels to the right of each diagram) show that up to the 850-millibar level a stream of northwesterlies was replaced by southeasterlies; yet this change in the air stream would not be

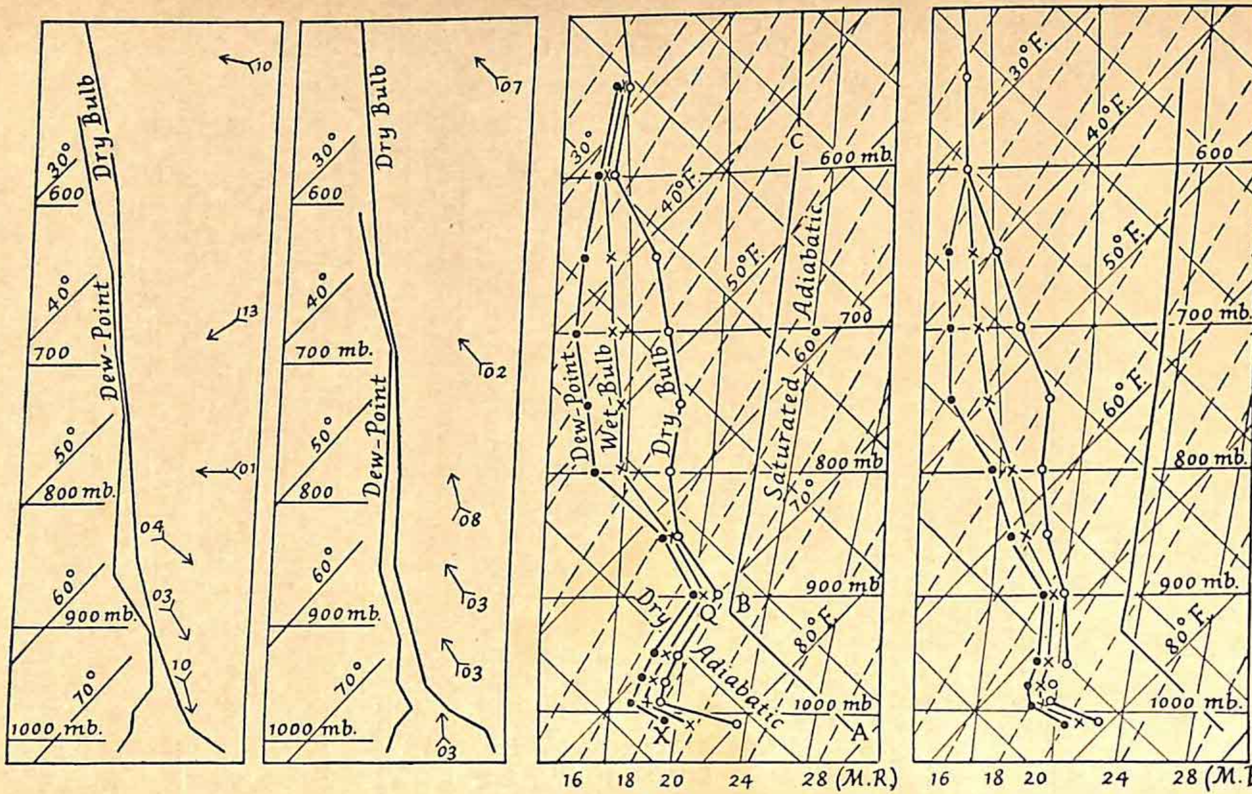


Fig. 16 Tephigrams—Seletar

(*a*) 0200 G.M.T., 5.11.51 (*b*) 0200 G.M.T., 6.11.51 (*c*) 0200 G.M.T., 24.1.51 (*d*) 0200 G.M.T., 25.1.51

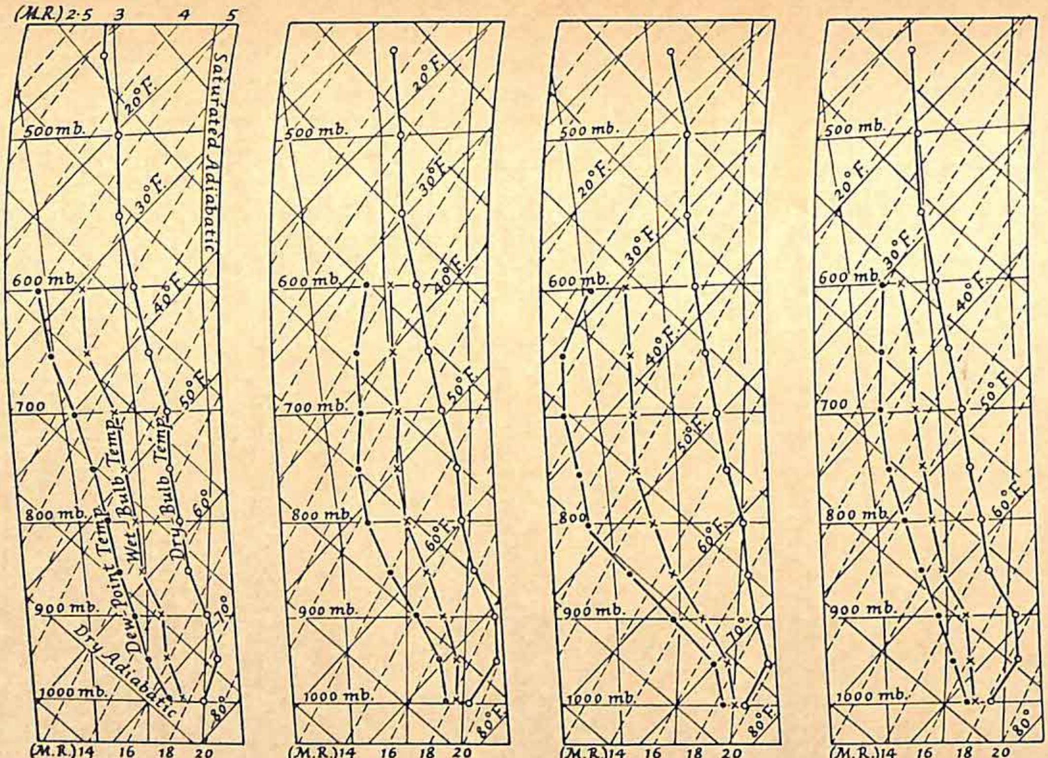


FIG. 17 Mean Tephigrams—Singapore

(*a*) January (*b*) March (*c*) July (*d*) December

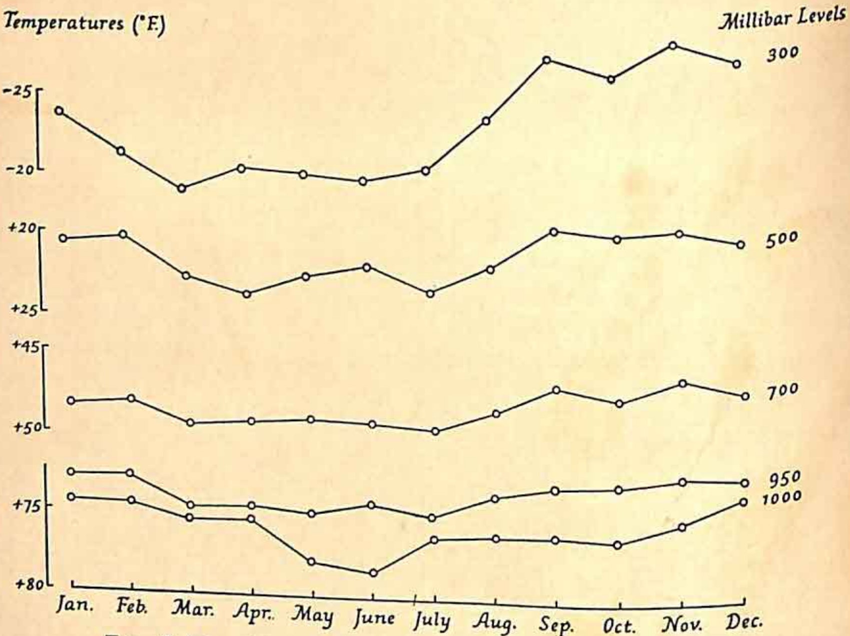


FIG. 18 Mean Monthly Temperatures at Singapore. (After John)

suspected from the tephigram, as lapse-rates and dew-point curves are practically the same on both days.

The second way in which the tephigram may be used in middle latitudes is for predicting the formation of cumuliform cloud and convectional rain by recognition of instability in the atmosphere. Such predictions are unreliable near the Equator, as, although the atmosphere is invariably moist and conditionally unstable, cumulonimbus forms on some days and not on others.

For example, Figs. 16*c* and *d* are plots of upper soundings at 0200 G.M.T. (0930 M.T.) on 24th and 25th January 1951. Since the general lapse-rate of temperature in each case is between the dry and saturated adiabatic rates, both the curves are conditionally unstable. The maximum surface temperature at Seletar on the 24th was 88° F. If the surface is warmed to this value (point A of Fig. 16*c*), the surface air will rise, cooling at the dry adiabatic rate (line AB). At point B it reaches saturation, since the dry adiabatic intersects the mixing-ratio line through X, the surface value of moisture content. Therefore, cloud should form with base at the 920-millibar level (i.e. between 2500 and 3000 ft.).* Since the

* The fall of pressure with height is about 1 millibar per 30 feet, but can be calculated more precisely from the tephigram.

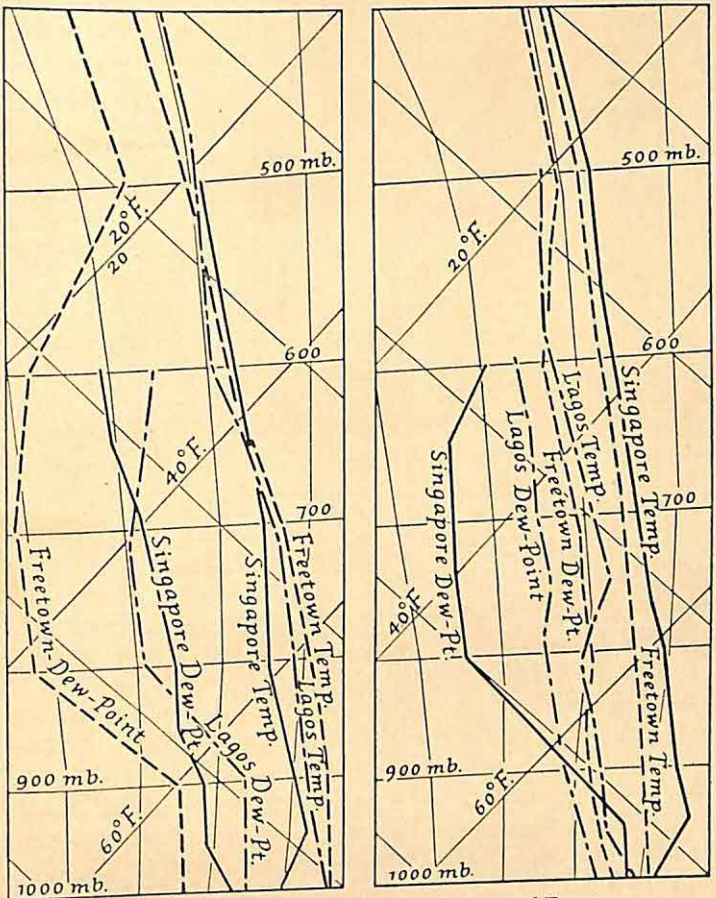


FIG. 19 Temperatures at Singapore, Lagos and Freetown
(*a*) Mean January Tephigram (*b*) Mean July Tephigram

rising air at B is warmer than its environment Q, it will continue to rise, but thereafter cools at the saturated adiabatic rate. At all points of its ascent, BC is warmer than its environment, and there is no hindrance to convection. Thus it is to be expected that, once currents are initiated in such an atmosphere, thick towering clouds will develop. A similar condition is found in the ascent of the 25th January 1951 (Fig. 16*d*). Despite the similarity, 2·29 inches of rain fell at Seletar on the 24th while rain was negligible on the 25th.

At the Equator there is little difference in lapse-rate throughout the year, as may be seen if the mean tephigrams for January, March, July and December (Figs. 17*a*, *b*, *c* and *d*) are compared.

Above the surface layer the mean lapse-rates in all months are conditionally unstable. The departures from these means are small, and therefore clouds of great vertical extent might be predicted daily. Much towering cloud does develop, but the amount over any unit of ground varies greatly from day to day without correlation with any changes on the tephigram.

4. Upper-air Mean Temperatures Over the Equator

Although the day-to-day variations in upper-air temperature and moisture content are small, there are some seasonal differences (Fig. 18). The most important is the fall of temperature at all levels from February to March, and the general rise of temperature from July to September, the annual range being greatest at very high levels.

Too few upper-air temperature soundings have been taken in equatorial latitudes to give a true picture of the horizontal distribution of temperature at various levels. However, monthly mean tephigrams are available[9] from Freetown and Lagos, which are located on the West African coast at latitudes $8\frac{1}{2}°$ and $6\frac{1}{2}°$ N. respectively. Comparing these means with those of Singapore, it appears probable that there is very little gradient of temperature at any level over the whole of the equatorial belt. In January (Fig. 19*a*) the Singapore temperatures at higher levels are practically identical with those of West Africa, and the greatest differences (at 800 millibars) are of the order of 4° F. In July (Fig. 19*b*) the Singapore temperatures at each level are within one to three degrees of those over Freetown, although temperatures at Lagos are somewhat lower.

CHAPTER IV

Formation of Fog and Cloud at Low Latitudes

Owing to frequent instability at low latitudes, it might be expected that cumuliform types would predominate and that stratiform cloud would be rare.

However, stratiform clouds are not unusual and their mode of formation merits some study.

1. Radiation Fog

The clear, calm nights required for the formation of radiation fog are very common in low latitudes and sufficient moisture is usually present, but fog formation is confined mostly to places inland. Fog is scarce on the coast, partly because a moderate wind normally blows seaward from the coast each night enabling ground cooling to be spread through a thick layer. The following figures compare the frequencies with which fog forms at coastal and inland stations:

Coastal Stations	*No. of Days of Fog per Year*
Labuan (North Borneo) . .	0
Singapore	0
Kota Bharu (Malaya) . .	9
Inland Stations	
Kuala Lumpur (Malaya) . .	52
Kuala Lipis (Malaya) . .	236
Temerloh (Malaya) . . .	78
Kluang (Malaya) . . .	86

As radiational cooling first occurs at the ground it is not uncommon to find that the ground temperature falls lower than that at 500 or 1000 ft., and so an inversion is formed simultaneously with the appearance of the fog. The inversion prohibits upward currents of air and maintains a flat top to the fog, which will not be dissipated until the increase of ground temperature by solar heating is sufficient to destroy the inversion, allowing the rising air to disperse the fog into the upper atmosphere. The length of time during which fog persists after sunrise depends on the strength of the inversion.

Radiation fogs generally dissipate about 9 a.m., but they may

persist inland until after 10 a.m., particularly where the presence of cloud (usually altostratus) at an upper level retards the warming of the ground (Plate I). Sometimes a slight wind increase after sunrise stirs up the lower air so that a greater thickness of air is cooled, causing the fog to thicken temporarily before heating becomes great enough for its destruction. These fogs are normally only a few feet thick, but occasionally may reach 50 ft. above ground.

As a rule, there is a greater frequency of radiation fogs in valleys. The sloping sides of valleys lose heat by nocturnal radiation, and the air thereby chilled tends to slide downwards because it increases in density due to cooling. It mingles with the moist air in the valley bottom, where the combination of low temperature and high humidity produces condensation, so that a fog layer eventually covers the valley. The effect is noticeable in the valleys of Malaya and Sumatra, where thick, persistent fogs frequently occur. An early morning view of valley-fogs in the neighbourhood of Kuala Lumpur is shown in Plate II.

Though fog is normally dispersed by up-currents following the destruction of the inversion, cases occur when the fog is lifted bodily upwards as a layer of stratus. Conditions for the two possibilities may be illustrated as follows:

(*a*) *Height*	*Evening Temp.*	*5 a.m. Temp.*	*10 a.m. Temp.*	(*b*) *Height*	*Evening Temp.*	*5 a.m. Temp.*	*10 a.m. Temp.*
feet	° *F.*	° *F.*	° *F.*	*feet*	° *F.*	° *F.*	° *F.*
3000	73	73	73	3000	73	73	73
2000	77	77	79	2000	77	77	77
1000	81	76	85	1000	81	76	76
ground	85	74	91	ground	85	74	82

In (*a*) ground temperature drops to 74° F. by 5 a.m., and a temperature inversion is formed between the ground and 2000 ft. Heating during the following morning destroys the inversion, and by 10 a.m. the lapse-rate has become unstable and any fog dispersed by up-currents.

In (*b*) the nocturnal fall of temperature is similar, but heating after sunrise does not influence the 1000 ft. level. Instability is set up between the ground and 1000 ft., allowing up-currents to carry the fog upwards. However, between 1000 and 2000 ft. a portion of the inversion still exists at 10 a.m., and the fog, having risen to the 1000 ft. level, must remain there as a sheet of stratus because the

inversion prohibits further upward movement. In the tropics, it is not common for such stratus sheets to persist because insolation destroys the inversion before mid-morning.

2. Sea Fog

As the air streams at low latitudes nearly always come from higher latitudes, it is rare for a warm stream to overlie a cooler sea, so that sea fog is practically unknown there. On the other hand, warm, moist currents of air from the tropics frequently pass into temperate regions and induce fog formation. Sea fog is common over the South China Sea during March and April, although negligible at other times. Pressure over Asia decreases during March, when the general northeasterly flow over the China Sea is occasionally interrupted by air streams of equatorial origin. Sea temperatures are then comparatively low, and fog, which forms readily in the equatorial air, may persist for several days at a time. It shows little tendency to break up during the day because insolation has not much immediate effect on sea temperatures, but it sometimes lifts to low stratus, particularly when carried to a coast-line.

3. Cloud Formation in Convection Currents

Most equatorial clouds form in currents of rising air, which cool by expansion. Strong upward currents are initiated by the differing capacity of types of ground to absorb solar heat: a surfaced road is warmed quicker than fields, and bare ground quicker than grass-covered land. Air travelling across a warm area is heated from below and begins to rise; thereafter ascent is under the control of the lapse-rate in the surrounding air. If that lapse-rate is steep, the currents may go to great heights; if inversions are present or the lapse-rate is slight, the currents will have a low upper limit.

If the lapse-rate is steep, the air rises and cools at the dry adiabatic lapse-rate until temperature reaches dew-point. Moisture then starts to condense out in a level cloud base. Above this 'condensation level,' the rising air cools off at the saturated adiabatic rate until, at some level, it has cooled to the same temperature as its surroundings. The cloud tops can be carried no higher than this level. Inversions are frequently found at some moderate height, and should they exist in the region of rising air, only fair-weather cumulus is formed (Plate III). This typifies a fine morning, the formation reaching maximum height about 2 p.m., and after sunset either dissipating or spreading out into a layer of stratocumulus.

The base of fair-weather cumulus is generally 2000 to 3000 ft. and the tops 4000 to 6000 ft. When the general air movement is sluggish

and conditions are favourable for convection, the atmosphere will probably contain both up-currents and down-currents. Brunt[10] quotes 6 m.p.h. and $1\frac{1}{2}$ m.p.h. as reasonable averages of the up-currents and down-draughts respectively. He states, furthermore, that in the absence of a variation of wind with height, convection cumuli are likely to be spaced so that the mean distances of adjacent clouds are $2\frac{1}{2}$ to 3 times the height of the tops of the cloud (Plate IV).

ABCD in Fig. 20*a* is the plot of temperature against height on a particular day when an inversion exists at BC. Ground heating raises the air at A until dry adiabatic cooling during ascent has

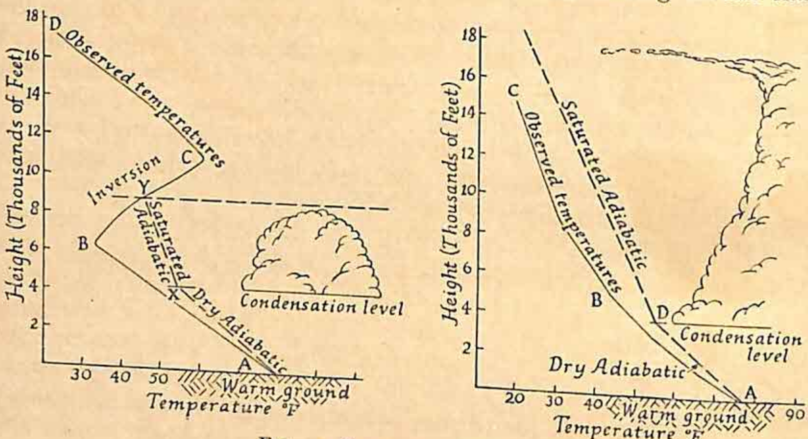


FIG. 20 Convectional Clouds

(*a*) Cumulus formed in Air rising beneath an Inversion

(*b*) Cumulonimbus; when up-currents practically unlimited

reduced its temperature to dew-point at X. Further ascent entails saturated adiabatic cooling until the rising air reaches the same temperature as the surrounding air at Y, and no further ascent is possible. All the cumulus clouds will have a common base at the X-level and tops at height Y.

If the lapse-rate is steep up to a considerable height, there is no stay to the upward growth of the cloud, and cumulus may tower to 10,000 or 20,000 ft. The ultimate stage of cumulonimbus may only be reached when the rising air attains equilibrium at a very great height, perhaps 20,000 or 50,000 ft.

The growth of a cumulonimbus is shown in Fig. 20*b*, where the lapse-rate ABC is steep throughout and the atmosphere is inversion-free. Ascent continues above the condensation level D, because the rising air is still warmer and lighter than the neighbouring atmo-

sphere, and cloud tops are carried to great heights before spreading to anvil form (Plate V). As all cumuliform or convectional clouds are built in up-currents, they are often accompanied by varying surface winds of air moving inward to the core of the cloud to compensate for up-draught. Clouds of great thickness—towering cumulus and cumulonimbus—frequently contain vigorous up-currents, whereas convectional clouds of small vertical extent—fair-weather cumulus—contain small up-currents. The mean values of the up-currents in clouds have been assessed by Ludlam[11] at 2 m.p.h. (1 metre per second) in small clouds, and over 10 m.p.h. (5 metres per second) in thick ones.

Over the land, cumuliform cloud develops gradually during the day and dissipates by sunset. Full development may be delayed until late in the afternoon and, more rarely, a cumulonimbus may continue growing after sunset. Cumuliform cloud grows with the steepening of lapse-rate when the earth becomes heated, and its cessation and final collapse follows an increasing stability when the ground becomes cooler near sunset.

If layers of stratus and altostratus are also present during the late afternoon, nocturnal radiation from the earth is restricted. This slows the fall of ground temperature, enabling instability to persist later than usual. When, because of the presence of a stratiform sheet, cumulus cloud remains active until sunset, it is possible that there may be out-going radiation from the cloud-tops after dark. This tends to produce lower temperatures aloft. Because the stratus or altostratus retards any temperature drop at lower levels, instability may temporarily be increased and the cumulonimbus develop further.

The diurnal variation of cumuliform development over the land is not the same as that over the sea, where there is a slight maximum of cloud growth during early morning and a minimum by day. Sea temperatures vary little from day to night, because solar heating of the sea surface is quickly spread through a great depth by wave motion. Early morning increase in cumuliform cloud over the open sea must be brought about by a steepening of lapse-rate through radiation from cloud tops.

The daily change of cumuliform cloud over coastal waters is extremely great and is brought about by the land breeze.[12] As temperatures fall over the land after sunset, the cooled air flows seaward, undercutting the off-shore air. The equatorial atmosphere is nearly always conditionally unstable, and the uplift which follows undercutting off the coast favours cumuliform growth.

The early morning maximum of cloud offshore has a marked

effect on precipitation at many coastal stations, because, if the prevailing upper wind is onshore, cumulonimbus then drifts landward. Thus coastal stations may have two maxima of cloud growth: the normal inland type occurring in the afternoon, and that created by shower clouds drifting ashore from coastal waters during the early morning.

The incidence of maximum cumuliform development and its accompanying maximum precipitation may vary considerably along a coast. Braak[13] investigated the question by observations made at Discovery Oostbank, a small reef between Borneo and Billiton, where maximum precipitation was between 5 a.m. and 2 p.m., and the minimum between 6 p.m. and midnight. He considered that the land breeze from Sumatra influenced this diurnal variation.

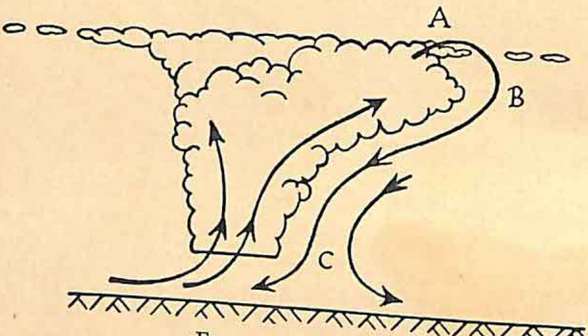


FIG. 21 Cold Squall

The extent to which early morning cumulonimbus develops over coastal waters partly depends on the curvature of the coast.[12] If it is concave, land breezes from its different parts tend to converge towards the centre of the concavity, undercutting and uplifting the air offshore. With a convex coast-line, the land breezes are divergent and the undercutting is less concentrated.

A squall of cool air frequently occurs during the passage of a tropical cumulonimbus cloud. The air composing this squall is believed to be a down-draught which has travelled through a cloud-free region of falling rain[14] (Fig. 21). Although the air leaving the cloud top (A) is saturated, much of its moisture has condensed into cloud following ascensional cooling. It then begins to descend (BC) and, as its temperature rises dry-adiabatically, the difference between temperature and dew-point increases. If this dry current passes through a region of falling rain, the water droplets will be evaporated into the current, entailing a fall in temperature.

PLATE I: FOG LAYER WITH ALTOSTRATUS ABOVE

PLATE II: VALLEY-FOG NEAR KUALA LUMPUR

PLATE III : FAIR-WEATHER CUMULUS

PLATE IV : SPACING OF INDIVIDUAL CUMULI

PLATE V : CUMULONIMBUS WITH TOWERING CUMULUS

PLATE VI : OROGRAPHIC CUMULUS OVER EASTERN SELANGOR

PLATE VII: ALTOCUMULUS CASTELLATUS

PLATE VIII: ALTOSTRATUS FORMED BY SPREADING CUMULONIMBUS TOPS

The direction in which a cumulonimbus moves is that of the main flow of air aloft, with an apparent correlation between the movement of such clouds and the direction of the winds in the layer of air between 8000 and 14,000 ft.[15] The speed of travel is probably much less than the wind speed; many columns remain practically stationary for long periods over their heated sources until the clouds are fully developed. Then, as the vigour of the up-currents decreases, the cumulonimbus may dissipate and, in doing so, drift down-wind—an effect frequently observed over the small scattered islands of the Rhiouw Archipelago. Individual cumulonimbus columns form over each island during the day, but, as insolation and convection decrease in the late afternoon, the clouds drift down-wind over the sea.

Over the large land masses, however, the columns and their attendant thunderstorms appear to drift with the wind before reaching maximum growth. This is not necessarily a bodily movement of the column; it could result from propagation, a succession of new thunderstorms forming downwind as the parent columns die. As the down-draught from a cumulonimbus is likely to travel with the general wind direction, it may undercut the air immediately to leeward of the original column, and, by doing so, create a new thunderstorm. Down-wind 3 miles away from an existing cumulonimbus is a place favourable for new cumuliform development.[16] Recent observations of thunderstorms by radar have shown that this mode of progression and propagation is common, and that, while individual columns are never long-lived, new ones are constantly being formed.

If velocities in the general wind stream are not great, the downflow from the top of a cumulonimbus spreads out evenly in all directions, causing the most likely place for the development of a new column to be between two existing ones. Byers[16] states that new development is most probable between two columns not more than 3 miles apart, where two cold outflows meet to displace the warm air upward.

4. Orographic Cloud

Because the equatorial atmosphere is usually moist and conditionally unstable, orographic cloud forms readily. But the winds are generally too light to enforce great uplift, so that both cumuliform and stratiform types are found (Plate VI).

Examples of orographic formations are plentiful in Southeast Asia. During the Indian Southwest Monsoon practically continuous layers of stratus, altostratus and nimbostratus at various levels line

the Thai and Burmese coasts from Phuket to Chittagong, possibly with protrusions of cumulonimbus at the coast during early morning and inland during afternoon. Stratiform layers also cover the east coasts of Indochina and Malaya during the Northeast Monsoon, accompanied by a tendency at inland locations for scattered cumuliform columns to rise through the strata about midday; offshore cumulonimbus often drifts to the coast in the early morning. Similar conditions are experienced on the northern coast of Borneo and Java during this monsoon.

5. Turbulence Cloud

Turbulence cloud is not common at low latitudes where, though the moisture content is high, winds are light and there is usually insufficient frictional stirring for the eddies to rise above condensation level. Wind speeds may sometimes be great enough in the monsoons to produce turbulence, and then rolls of stratocumulus occur over exposed coasts.

6. Cloud Formed by Shear

A formation similar to turbulent cloud may derive from a different cause. When there is an increase of wind strength with height or a marked change of wind direction within a short range of altitude, there is said to be 'shear' through the layer in which the differences occur. The variations of wind speed or direction cause eddies in the shear layer, and rolls or globules of cloud may be produced.

Brunt[10] describes the effects of shear thus:

(1) When there is a rapid increase of wind with height, the clouds form in long rolls with ascending motion over the whole of the cloud base.

(2) When there is only a moderate variation of wind with height, cloud develops mostly in long rolls transverse to the direction of shear; these are often distorted by slight local wind variations producing other shorter rolls.

(3) Clouds which are in the form of small globules arranged in roughly parallel lines indicate a moderately small shear of wind with height, in conjunction with a fairly marked instability within the layer.

(4) Clouds which show two systems of wavelike patterns at right angles to one another indicate a low rate of shear of wind and a small degree of instability.

No satisfactory explanation has yet been produced for the formation of those extensive cloud sheets observed in the monsoons of

Southeast Asia. Indistinct rolls are sometimes evident in them, so that some of these formations are probably due to shear within a layer. At other times the layers are thick and uniform, and a different cause must be sought. It is likely that, in the latter case, a gradual convergence is occurring over a great horizontal area, leading to a slow general ascent of air within the mass.

It is not certain what induces this convergence in the monsoon stream, but it evidently exists since, during the Southwest Monsoon, velocities in the south-southwesterlies on the Tenasserim coast are frequently much lighter than in the west-southwesterlies crossing Ceylon. Probably the effect is related to orography; the stream turns bodily in the eastern Bay of Bengal to run northward parallel to the ranges of Burma.

Cloud is usually dissipated by subsidence: near the edges it breaks into globules, which disappear in the heating consequent to the subsidence.

Instability sometimes develops within a cloud sheet, either through radiation cooling of the tops and heating at the base, or by subsidence of the air in which the clouds are suspended.[10] In the first case, the lapse-rate is steepened, and cumuliform columns may emerge from the layer. In the second, the air below the cloud is heated dry-adiabatically during descent, but the air within it may warm at the lesser saturated adiabatic rate. This differential brings about an unstable lapse-rate within the cloud and conditions favourable for cumuliform development. A common result of increasing instability is the development of altocumulus castellatus (Plate VII).

7. Spreading Tops of Cumulus Clouds

Many stratiform sheets are formed by a spreading of the tops of clouds which are cumuliform in origin. The upward currents in a cumulonimbus cloud, on meeting a stable layer, spread out in anvil tops at altitudes between 18,000 and 50,000 ft. (In temperate regions, the tops are much lower.) Air currents, which have ascended in the cumulonimbus, flow horizontally to join the main atmospheric stream, carrying with them cloud particles to compose a stratiform sheet down-wind from the original column (Plate VIII).

Normally the sheets are of cirrus, sometimes of altostratus at various levels. The latter are formed either through horizontal outflow at altitudes less than the cloud tops, or by the spreading tops from towering cumulus which have not reached cirrus altitudes.

The incidence of altostratus sheets formed by the spreading of cumuliform tops is low during the middle of the day when convec-

tion is at its peak, because of the down-currents in the neighbourhood of each cumulonimbus. In the late afternoon when convection usually decreases, the down-draughts also decrease and conditions favour the forming of altostratus and nimbostratus. The layers formed thus are frequently of considerable depth and may produce rain. They may persist all night; they become broken if instability develops during the night, favouring the formation of new cumuliform columns and dissipation of the sheets between them. The instability may come through radiation from the tops of the sheet, or by the steepening of the lapse-rate aloft following subsidence consequent to the radiational cooling of the lowest layer of the atmosphere.

8. Cloud Formed at the Boundary of Two Air Streams

Cloud concentrations are particularly great at the boundary of two different currents of air, where winds converge and air ascends, encouraging cumuliform development. Sheets of altostratus and cirrostratus formed in the spreading of these cumuliform clouds are persistent because, as these convectional columns are not formed by local heating, they are not necessarily accompanied by down draughts. Furthermore, where the opposing currents meet, shear may develop in various strata, together with bodily uplift of portions of the currents. These effects also contribute to the formation of layer-clouds near the boundary. The result is that, at a boundary with convergence, conditions favour the development of both cumuliform and stratiform clouds.

CHAPTER V

Strong Convection and Precipitation

1. Thunderstorms and Precipitation

Cumulonimbus clouds are associated with thunderstorms and heavy, localised rain. Within them, to the danger of aircraft, there is much turbulence because of the convection which develops regions of contrasting up-currents and down-draughts. Recent observations[17] by aircraft flying through cumulonimbus clouds at various heights between 4000 and 26,000 ft. have shown that turbulence is least near 4000 ft. (which is nearest the base of the cloud) and increases up to a height which is 10,000 ft. below the maximum height reached by each cumuliform column. Jones[18] states this in another form, that turbulence within a cumulonimbus cloud increases gradually from the base upwards until a height of a little less than half the total vertical extent of the cloud is reached. Above this level, turbulence remains nearly the same for a considerable height.

It is generally accepted that condensation in the atmosphere takes place on hygroscopic nuclei. It has been stated[19] also that, when ice particles are co-existent with water particles in a cloud, rain of appreciable intensity develops. However, the presence of ice crystals is not essential in the production of rain, as recent observations[11, 20, 21] in the tropics indicate that rain frequently falls from clouds of which the tops are below freezing level.

Recent studies of cloud formation have included investigation into the function of the nuclei of condensation. Experiments have included introducing substances, such as crushed 'dry ice' (solid carbon dioxide), silver iodide and volcanic dust, into an existing cumulus or layer of stratocumulus. The results of 'seeding' of clouds in this way have been that cumuliform columns with rain have developed where conditions were apparently favourable, and that simultaneously cloud has dissipated in areas where horizontal divergence already existed.[22]

Thunderstorms are very common in low latitudes, and in most places they are a daily occurrence during certain seasons. Inland maximum frequency is in the afternoon, while over the open sea it is at night or early morning. Coastal areas commonly have two maxima—one in the afternoon through local heating and one in

the early morning associated with cumulonimbus development off-shore. At Penang thunder is heard on an average of 204 days per year. It is heard during daylight hours on 124 days and at night-time on 152 days. Over most of England there are only about 10 to 15 days of thunder each year.

Thick altostratus and nimbostratus in the tropics sometimes contain strong enough up-currents to produce moderate or heavy falls of rain, and these conditions are most often experienced when the layer cloud is orographic or associated with the spreading currents of a cumulonimbus—where the flow in the layer may have an upward component. Occasionally at low latitudes, when a sheet of altostratus extends outwards from a large cumulonimbus, a separate cumuliform column grows out of the altostratus sheet.

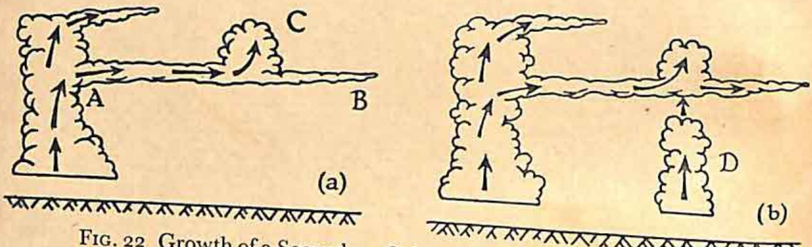


FIG. 22 Growth of a Secondary Column from the Altostratus Sheet

An example of cumulus growing from an existing sheet of altostratus was observed during June 1951. The parent column was a cumulonimbus which had developed in the early afternoon over Pulau Batam (an island 15 miles south of Singapore), and from this column a sheet of altostratus about 3 miles wide and 1000 ft. thick stretched down-wind for 15 miles (AB of Fig. 22*a*). A secondary column (C of Fig. 22*a*) developed with base in the altostratus and, as the second column was over the sea, it appears probable that it was fed by currents running through the altostratus and initially supplied by the parent column. An hour later, the up-currents of the second column induced the up-growth of a third cumulus (D of Fig. 22*b*) immediately below the second column. Finally, the tops of the new cumulus penetrated the altostratus and formed, with the secondary column, one continuous towering cumulus.

2. Intense Convection*

Tornadoes and waterspouts are occasionally found at low latitudes. Since they are of a convectional nature, instability is a

* With acknowledgment to Gordon.[23]

condition for forming them, and this may often be due to insolation over land. A steep gradient in the air's moisture content also favours convection, because water vapour is less dense than dry air. Upward motion is apparently initiated when some small portion of air near the earth's surface becomes warmer and moister than the surrounding air. The rising air pursues a spiral motion upwards, the rotation being frequently (but not always) clockwise in the Southern Hemisphere and counter-clockwise in the Northern Hemisphere. Moisture usually condenses as a column of cloud in the rising current, and latent heat thus released supplies additional energy for the circulation.

Tornadoes and waterspouts are frequently associated with cumulonimbus clouds, and it is usually considered[24] that convection

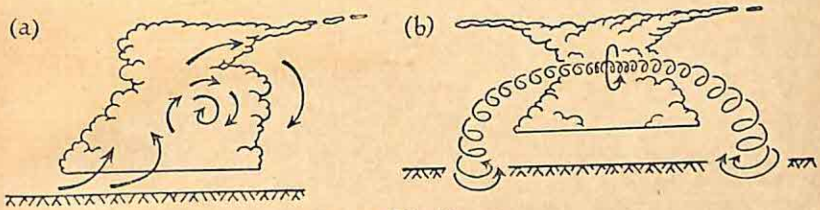


FIG. 23 The Vortex
(*a*) Cross-section of a vortex (*b*) The two vortices

into the base of a cumulonimbus initiates the vortex. Wegener,[25] while agreeing that its development is associated with a cumulonimbus cloud, considers that the vortex initially lies along a horizontal axis in the forward part of the cloud and across its direction of travel (as in the cross-section of Fig. 23*a*). Visible waterspouts are the ends of the vortex, which has two spouts bending downwards as they emerge on either side of the cloud. These two spouts revolve in different senses at ground level (Fig. 23*b*). Supporting this theory, pairs of spouts associated with single cumulonimbus columns have been observed in the Java Sea.

From the ground, tornadoes appear as funnels of cloud, sometimes curved and normally capped by thick cumuliform cloud. Their frequencies are low at very low latitudes, and greatest in North America and Australia. They are highly destructive, the damage being caused by strong winds (at times exceeding 200 m.p.h.) and by explosive effects of the extremely low pressure within them.

Waterspouts are usually first observed in an inverted cone of cloud emerging from the towering cumulus or cumulonimbus. Its lower and narrower end wavers as it grows downward, and on

reaching low levels agitates the surface of the sea as water is lifted bodily to form a continuous funnel stretching from sea to cloud. This funnel narrows to a waist at middle heights. The spout may travel down-wind erratically at speeds of 10 to 20 miles per hour.

Decay sets in at low levels when the funnel narrows following a decrease of winds, and the narrowing and dissipation progress upwards until all sign is lost.

Gordon[23] estimates that although short-lived waterspouts of 200 ft. vertical length are common and the longest observed spout was 5000 ft., the average length is between 1000 and 2000 ft. Widths vary from a few feet to 800 ft. in diameter, and two extreme cases are quoted,[23] one where the vertical length was 100 ft. and diameter 700 ft., and another whose length was 1050 ft. for a diameter of only 3 ft. Spouts in equatorial waters are generally 500 to 2000 ft. in length and about 50 to 100 ft. in diameter.

3. Equatorial Rainfall Types

Most rainfall in low latitudes is associated with cumulonimbus clouds, so that the diurnal variation of rainfall depends largely on that of the clouds. Maximum precipitation (and thunderstorm frequency) occurs by day over land, and at night or early morning over the sea. Falls from altostratus or nimbostratus contribute to the total rainfall, but without marked diurnal variation.

Rainfall in the equatorial region is of four types:

(*a*) Rain at the boundary of two large-scale air streams: Where they meet, each has a component of velocity towards the common boundary, the resulting inflow being relieved by up-currents favourable for the development of cloud and precipitation. Such cloud is predominantly cumuliform, though stratiform layers are common. This type of precipitation is found at the onset of the Indian Southwest Monsoon in Burma, where some of the heaviest rains accompany the arrival of air from the southwest.

(*b*) Orographic Rains: When a stream of air is flowing over land, maximum precipitation within the stream takes place on the exposed coast. Orographic cloud is typically stratiform, but cumuliform is likely at certain times of the day.

(*c*) Convectional Rains: These are associated with individual convection columns within a stream, and wholly subject to the diurnal régimes peculiar to land, sea and coastal regions. Rain may fall from the altostratus spreading from the chief cumuliform columns.

(*d*) Rain from convergence within a single large-scale air stream: Convergence within a stream may favour the formation of cumulo-

nimbus and altostratus (Chapter X), leading to precipitation primarily from the cumulonimbus, but also extensively from the accompanying altostratus sheet.

Although rainfall types may be classified according to their formation, precipitation at any one time may derive from a combination of types. Some of the heaviest rain of the Indian Southwest Monsoon occurs from type (*a*), but the amount is far less over the Bay of Bengal than it is on the Burmese coast, where orographic uplift (type (*b*)) also comes into play.

CHAPTER VI

Rainfall Types of Southeast Asia

1. Factors Contributing to Rainfall

Maps showing the average rainfall for each month of the year do not throw much light on key features regarding the distribution of rain, such as the variation of its frequency and intensity with time of day and the relation of this variation to locality. Nor do they give a true indication of the monthly rainfalls which occur most commonly. Other interesting features which rarely appear on rainfall maps are the range of average falls from the wettest to the driest months, and the proportion between the average falls of the driest and the wettest months—which is of special interest in ecology.

A study of Table 4 shows that the mean annual rainfall of Southeast Asia is rather greater in amount than that of temperate latitudes. Annual rainfall varies from 50 inches in parts of Thailand and Cambodia to over 184 inches at Akyab in Burma, whereas annual falls in the British Isles are mainly from 30 to 60 inches, except in parts of the west where the average may exceed 150 inches. Another fact is the great seasonal range of rainfall in some tropical areas compared with others and compared with most temperate regions. At Akyab the range of the monthly mean values (the difference between the mean rainfalls of the wettest and the driest months) is 47·42 inches, at Singapore 3·48 inches and at London 1·42 inches.

These features are related to the dissimilarity of the factors contributing to the rainfall in tropical and temperate regions. Most falls at higher latitudes occur during the passage of disturbances covering a large horizontal area, and are distributed to correlate with orographic detail. Large disturbances of that kind are abnormal in tropical regions.

The greatest rainfalls of the Equator are often associated with the strong seasonal currents of monsoons and with exposed coasts. Burma's heavy rainfall in July is when the Indian Southwest Monsoon is strong there. Similarly, the Northeast Monsoon brings heavy rain to Indochina by October, to Kelantan and Trengganu by November, and to Northern Java by January (Table 4). In all these cases relief is one of the chief contributory factors.

At many places the greatest falls are when the monsoon begins. Braak[26] states, 'As a rule [in Sumatra] the monsoon change in the second part of the year is the principal rainy season. . . .' The onset of a monsoon involves the replacement of one current of air by another coming from a different direction, and along their boundary conditions favour the large-scale formation of rain. The advance of the boundary is slow so that the rain it causes in one locality may persist for a long time. For instance, the Northeast Monsoon moves fairly uniformly across the China Sea, but once south of latitude 5° N., its average speed decreases and at times its boundary may become stationary or even temporarily retreat. The rainfall accompanying the boundary may then persist for several days and add substantially to the total rainfall.

Linear disturbances wholly within an air stream may move down-wind and bring rain; they are not frequent in Southeast Asia though their part in the total rainfall must not be overlooked.

An isolated convectional shower in the tropics may produce a heavy fall of rain over a small area. The conditions for showers are nearly always present at low latitudes, except to the lee of mountain ranges where the drying-out of a steady monsoon current decreases the incidence of showers. Thus an exposed locality like Singapore has monthly rainfalls varying only from 6·66 inches in July to 10·14 inches in December, whereas at Kota Bharu they vary from 26·25 inches in December (when it is exposed to the Northeast Monsoon) to 5·52 inches in July, when southwesterlies cover Malaya (Table 4).

2. Regional Distribution of Rainfall

The mean monthly rainfall maps (Figs. 24 to 27) and mean annual rainfall map (Fig. 28) have been compiled from a number of sources [27-32] to illustrate the seasonal march of rainfall over Southeast Asia. Their implications in parts of the region are as follows:

Indochina

The Northeast Monsoon reaches Indochina about October, and its incidence is evident in the mean monthly rainfall at Quangtri —August 3·94 inches, September 15·71 inches and October 26·42 inches (Table 4). Heavy rain occurs on all the northeast coast during October. Leeward of the coastal ranges rainfall is as low as 4 inches, while farther south in Cambodia and Cochin-China, where shelter is less, the monthly mean is 8 to 16 inches. September and October are the wettest months in Cambodia. Orographic rainfall remains high (about 16 inches) on the northern Indochina

Kuala Lumpur .	6·77	6·25	9·25	10·76	8·40	4·97	3·97	6·24	7·36	11·05	10·10	9·28	94·40
Mersing . .	14·74	8·08	8·26	5·47	6·36	5·88	6·43	6·82	6·13	9·40	13·53	20·23	111·33
Singapore . .	9·90	6·85	7·63	7·39	6·80	6·78	6·66	7·68	6·99	8·20	9·97	10·14	94·99
Sumatra													
Sabang . .	12·33	5·91	5·47	3·74	6·42	4·10	3·90	4·33	6·69	7·60	9·84	13·98	84·31
Medan . .	5·67	3·31	4·21	5·24	6·85	5·16	5·24	6·81	8·43	10·55	9·41	8·47	79·35
Padang . .	13·86	10·12	12·16	14·48	12·80	11·70	10·52	13·74	16·18	20·08	20·47	19·21	175·32
Palembang . .	11·26	9·65	12·25	11·18	7·17	4·76	3·82	4·02	4·33	8·03	11·14	12·64	100·25
Benkoelen . .	11·97	10·63	11·50	11·50	9·01	7·37	6·89	8·11	9·21	14·05	15·51	14·13	129·88
Borneo													
Pontianak . .	10·91	8·19	9·53	10·94	11·10	8·74	6·46	8·03	8·97	14·37	15·27	12·68	125·19
Balikpapan . .	7·87	6·85	9·09	8·23	9·13	7·60	7·13	6·38	5·51	5·24	6·61	8·15	87·79
Java													
Djakarta . .	11·81	11·77	8·27	5·79	4·45	3·78	2·48	1·65	2·60	4·37	5·59	8·03	70·59
Soerabaya . .	12·33	11·50	10·52	7·37	4·33	3·46	1·85	0·51	0·51	1·49	4·64	9·80	68·31
Jogjakarta . .	13·90	13·19	12·25	8·27	4·96	3·46	1·61	0·95	1·22	3·70	8·97	13·39	75·87
New York . .	3·76	2·94	4·17	3·57	3·31	3·89	3·35	4·14	4·79	3·11	2·62	2·67	42·32
London . . .	2·19	1·45	1·61	2·04	1·89	1·57	1·93	2·09	2·04	2·60	2·87	1·81	24·09

Table 4

Mean Monthly Rainfall (Inches)

	J.	F.	M.	A.	M.	J.	J.	A.	S.	O.	N.	D.	Year
Burma													
Akyab . . .	0·00	0·08	0·30	0·84	18·97	37·86	47·42	37·01	26·31	11·49	3·00	0·91	184·19
Mergui . .	1·13	1·90	2·33	5·52	15·19	29·89	32·70	29·06	23·99	11·57	4·28	1·30	158·86
Rangoon . .	0·04	0·28	0·09	4·00	12·20	21·41	23·25	20·21	17·21	7·86	3·24	0·97	110·74
Thailand													
Bangkok . .	0·03	0·78	1·41	3·11	4·72	6·37	5·85	5·55	11·11	7·49	2·16	0·52	49·10
Indochina													
Quangtri . .	6·03	1·22	1·46	1·34	2·76	2·64	4·84	3·94	15·71	26·42	15·99	7·21	89·56
Cambodia													
Phnom-penh .	0·39	0·32	0·75	2·40	5·40	5·43	5·75	5·40	9·29	8·90	5·12	1·42	50·57
Malaya													
Penang . .	3·48	2·99	5·77	8·21	10·33	7·39	6·57	10·84	14·14	16·63	12·01	5·87	103·95
Kota Bharu . .	10·19	5·46	7·00	4·61	6·41	6·09	5·52	6·59	8·71	11·96	24·04	26·25	122·83

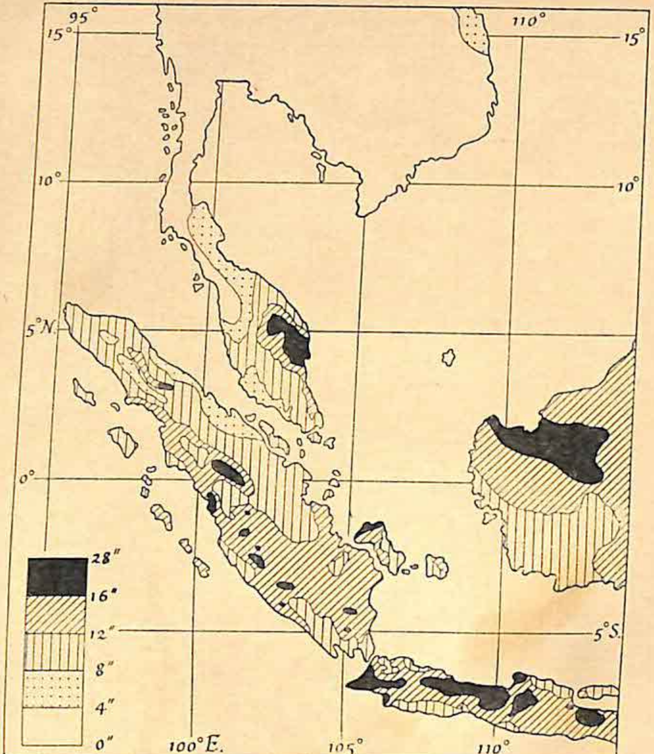


FIG. 24 Mean Monthly Rainfall in January—Southeast Asia

FIG. 25 Mean Monthly Rainfall in April—Southeast Asia

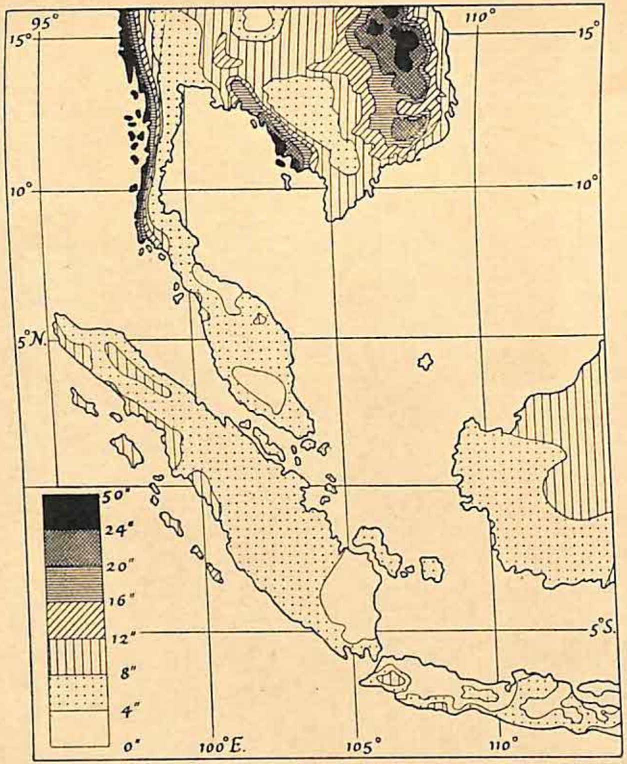


Fig. 26 Mean Monthly Rainfall in July—Southeast Asia

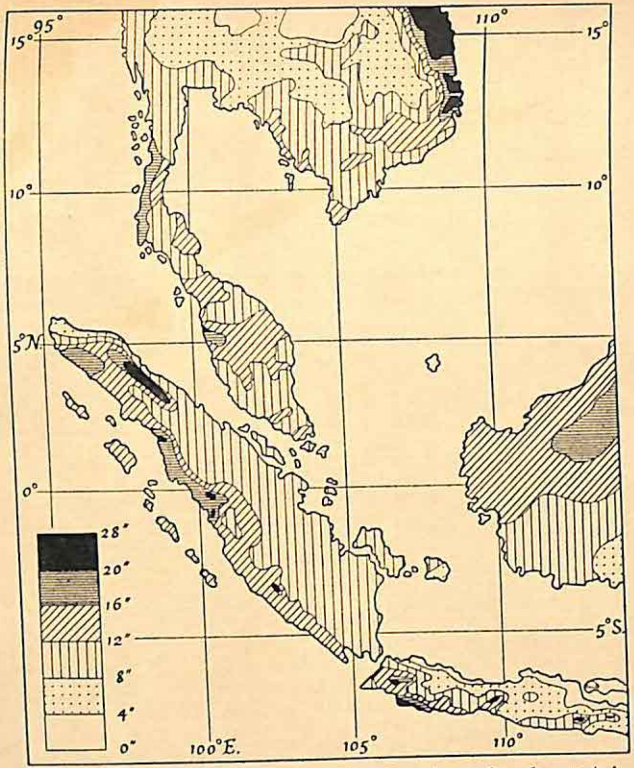


Fig. 27 Mean Monthly Rainfall in October—Southeast Asia

coast during November, but falls are then decreasing in the sheltered regions to the southwest.

Falls diminish generally during December and January, and by February are negligible over the whole area. Before the end of April the orientation of the current has changed, and the Northeast Trades enter Indochina as southerlies. Although the coastal belt is still nearly rain-free, precipitation increases on the ranges, becoming more marked in May as the leading boundary of the Indian Southwest Monsoon moves across the country, so that at Phnompenh the rainfall increases from 2·40 inches in April to 5·40 inches in May. Owing to sheltering, there is less increase on the northeastern coast.

From May to late September, the westerlies of the Southwest Monsoon cover the entire country, bringing rainfalls of 20 inches in exposed coastal portions of Cambodia and on the western slopes of the northeastern ranges. The central region remains reasonably dry, averaging about 4 inches per month.

From July to October there are about 20 to 25 rain-days* per month in the east and about 15 to 20 elsewhere. Frequencies thereafter slowly decrease, and from December to March there are less than 5 rain-days per month. After March rainfall frequency gradually increases—at first from showers, when instability builds up after the Northeast Monsoon has ended, and later from the Southwest Monsoon.

Thailand Above Latitude 8° N.

Heavy rainfall does not occur in Thailand at any time, because the country is sheltered from the Southwest Monsoon by the Tenasserim Ranges and from the Northeast Monsoon by the relief of Indochina. Thailand's mean annual rainfall ranges only from about 40 to 60 inches. The monthly total is highest (about 8 to 12 inches in places) during September and October when the Northeast Monsoon is advancing; for some months afterwards precipitation is negligible (less than 1 inch), because the monsoon current loses moisture while crossing the high northeast.

In June and July, maxima up to 16 inches occur on the coast east of Bangkok because, although the Indian Southwest Monsoon is depositing heavily over southern Tenasserim, a moist current still reaches the South Thailand coast. Rain-days are most numerous (10 to 15) in September; July, August and October normally have about 7 days each, after which the monthly frequency is less than 5.

* Rain-day—a period of twenty-four hours in which 0·01 of an inch or more of rain is recorded.

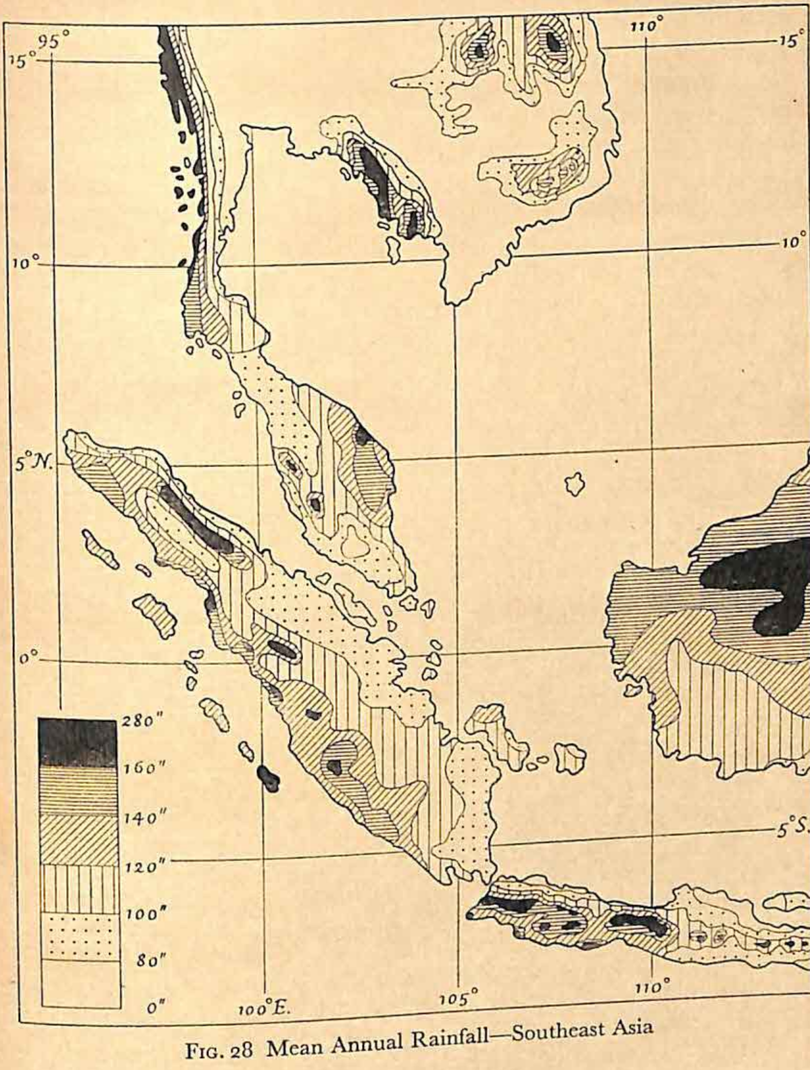


FIG. 28 Mean Annual Rainfall—Southeast Asia

Tenasserim

The Indian Southwest Monsoon comes over Tenasserim and all Burma during May. With its onset, heavy rain occurs along the ranges and western coast totalling from 40 to 50 inches in the far north to 10 inches south of Phuket. After May, falls slowly decrease until November, when the Northeast Monsoon is already established; then relief reduces rainfall to about 3 inches per month. A further decrease follows, monthly totals of less than 1 inch being

common until April, when convective showers (associated with the weakening of the Northeast Monsoon) produce averages from 1 to 5 inches.

There are 25 rain-days per month in the north and 15 in the south from June to August. The frequency lessens thereafter, becoming less than 5 per month from December until May.

Malaya and Southern Thailand

Stewart[33] recognises three distinct rainfall regions in Malaya. The first is along the east coast, extending about 30 miles inland; the second is a small area of the west coast about Malacca, stretching from Port Dickson to Batu Pahat; the third consists of all inland areas and the rest of the west coast.

The east coast derives its maximum rainfall from the Northeast Monsoon. Heavy monthly totals (20 to 25 inches) occur there at the onset of the Monsoon in November, and relief prolongs the wet season throughout December. Rainfall slackens during January, while from February to September monthly means are less than 8 inches. An increase to 16 inches in October precedes the new cycle.

The range of monthly means of rainfall around Malacca is not great—from a minimum of 3 inches in February to a maximum of 10 inches in October and November, which relates partly to the leading boundary of the Northeast Monsoon and partly to showers. The remainder of the year has rainfalls attributable to showers. Rainfall from June to September is low, possibly related to sheltering from the Southwest Monsoon by Sumatran relief. The unique régime of the monthly totals during this period will be discussed later in this chapter. The contribution from afternoon showers is small, and most falls are from heavy showers formed in the Straits of Malacca during early morning (Fig. 35).

Central Malaya and the rest of the west coast are sheltered from the Northeast Monsoon by the ranges of Eastern Malaya, and from the Southwest Monsoon by those of Sumatra. Consequently, highest rainfalls occur between the monsoons when, owing to slack currents, no föhn effects lessen the shower activity. Monthly rainfalls are 3 to 8 inches, rising to maxima of 11 inches in April and October. Near Penang, the earlier maximum is delayed until May when the leading boundary of the Southwest Monsoon arrives.

Thus rainfall in Malaya is a synthesis of boundary rain, of orographic rain in the monsoons and of showers. The contribution of the monsoons is best realised by comparing the monthly changes in any eastern portion of Malaya and those at Singapore. As

orographic influences are so small at Singapore, much of its rainfall is from showers which, though haphazard each month, result in an annual range of monthly means which is small.

Annual rainfall varies from place to place in Malaya between 80 and 120 inches, no one place being extremely dry or wet: the driest is Jelebu (65 inches per year) and the wettest recorded is near Taiping (232 inches).[34] Rain-days are fairly evenly distributed throughout the year, and their maxima coincide with highest mean monthly rainfalls. Thus in wet periods there are 20 to 25 rain-days a month in central and Eastern Malaya, but at no period and in no locality are rain-days fewer than 10 per month.

Sumatra

The rainfall of Sumatra is described by Braak[26] as follows: 'Generally speaking rainfall is abundant and fairly equally distributed over the year. As a rule the monsoon-change in the second part of the year is the principal rainy season, a smaller rainfall maximum occurring in the first monsoon-change, two relatively dry periods lying between the two maxima. In relatively small regions, where January and February are the rainiest months, the monsoon-change rains are joined with the monsoon rains into one rainy period, and in that case there is only one relatively dry season. Such monsoon maxima of rainfall occur in the southeast part of Sumatra, which is in this respect a region of transition with regard to Java and the Minor Soenda Islands.'

The mean annual rainfall is less than 100 inches in eastern areas and a small part of Northern Sumatra. Annual rainfall on the west coast and among the ranges is 120 to 140 inches, but in several regions falls are more than 160 inches annually. During May, June and July, when the Southwest Monsoon is established farther north, Sumatra is covered by a light southwesterly to southerly stream, the southernmost fringe of the monsoon. Rainfall is then about 5 inches monthly over Eastern Sumatra, and only 8 to 12 inches west of the ranges. A slight general increase occurs during August, when greater variability in the wind directions promotes showers, while from September onwards monthly rainfall is 8 to 12 inches on the east coast and over 12 to the west.

Rainfalls increase in December and January with the arrival of the Northeast Monsoon. Monthly rainfall is 8 to 16 inches then, though it reaches 20 inches in parts of the high country. There is a decrease during February, but in March and April rainfalls increase in the west with the beginning of the southwesterlies, which later in the year move north to form the Southwest Monsoon.

From March to May there are about 15 rain-days per month in the east and 25 in the west; from June to September they are fewer. A monthly total of more than 20 rain-days is common during the period October–January, while February rarely has as many as 15 per month anywhere.

Java

Regarding Java, Braak[35] says 'the principal difference between the climate of Java and Sumatra is the much sharper distinction between a dry and a wet season. This distinction is well developed in West Java, but much stronger in East Java and parts of Central Java, the greater part of which is practically rainless during several months.' During the dry season from May until October, Java lies in the Southeast Trades either from the Southwest Pacific or from the Australian continent. Then, monthly rainfalls are less than 5 inches except in the ranges, and in the east become negligible during August and September.

There are slight increases during November due to frequent showers. With the arrival of the Northeast Monsoon in December, monthly rainfall exceeds 16 inches over much of the country exposed to the north; elsewhere it is between 8 and 16 inches. Even after the monsoon boundary has passed to the south of Java during January and February, these high monthly averages are maintained by orographic rainfall in the northerlies. Although the Northeast Monsoon ceases before March, there is sufficient shower activity during March and April to maintain a monthly rainfall of 8 to 12 inches over much of the island.

During January and February there are about 20 rain-days per month to the north and about 15 to the south. A decrease of frequencies is apparent from April onwards, and the dry months July to September mostly have less than 5 rain-days per month. October sees a slow increase in frequencies, and by November most places have 15 to 20 days of rain per month.

Borneo

There are insufficient observations of rainfall over much of Borneo to provide detail of conditions there, though the broad trends may be seen. The onset of the Northeast Monsoon by December is marked by mean monthly falls up to 20 inches north of the ranges, while the sheltered southwestern region has less than 12 inches. Orography helps to maintain similar conditions during February, after which rain is of the shower type, and the total be-

comes generally less than 12 inches by April, except in the central ranges.

Showery conditions continue until June, when the Southwest Pacific Trades cross all Borneo, and rainfall drops to less than 8 inches over its southern half. The slackening Southeast Trades in October bring showers with increasing monthly totals and, until the arrival of the Northeast Monsoon, rainfall remains above 12 inches in the north and below 12 inches in the south.

3. Individual Observations and the Rainfall Mean

In making use of mean values of rainfall, it is necessary to exercise care. The annual rainfall and monthly mean rainfalls are each obtained from many observations varying considerably one from another. Actual conditions in a certain month may be different from the mean, particularly in the tropics where the range of the monthly totals is great.

Rainfall observations have been maintained at Taiping, Malaya, since January 1888. During the 624 months up to the end of 1939, there was a great range between the smallest and the largest of the monthly totals. The driest month was July 1929, with no rainfall, and the wettest was October 1902, with 44·45 inches. The mean rainfall for all the months of the year was 13·80 inches, that for July 6·51 inches and for October 20·50 inches. These means show that July is generally a dry month and October wet, but they inadequately describe conditions which can occur in a particular month.

We may investigate the distribution by dividing all the monthly totals into classes and determining the number of months during which the rainfall falls within each class. Thus rainfall was less than 1 inch on two of the 624 months, 1 to 2 inches during six of the

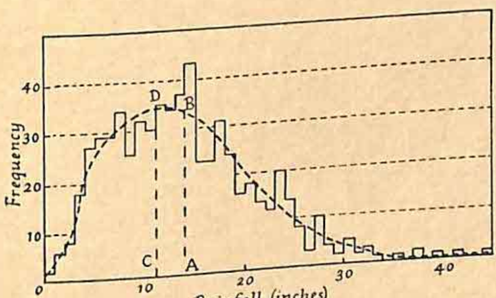


FIG. 29 Analysis of Monthly Rainfall at Taiping, 1888–1939

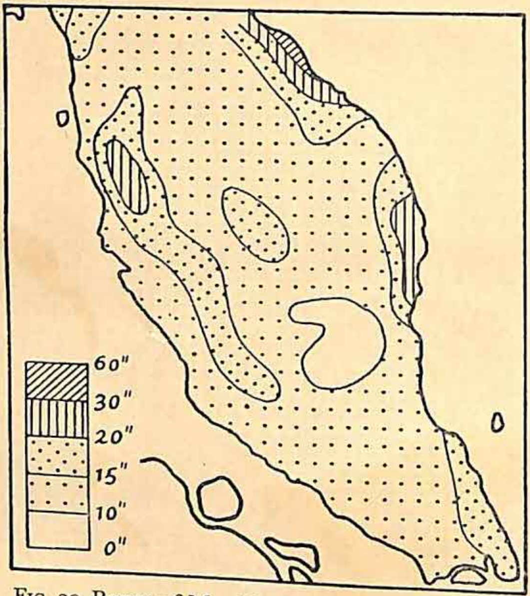


FIG. 30 Range of Monthly Rainfall, April—Malaya

months, and 2 to 3 inches during eight of the months, and so forth. The numbers in each class are plotted in Fig. 29, and a smooth curve drawn through the values. The mean value of all the months, 13·80 inches, is also drawn (line AB).

From Fig. 29 it appears that the mean does not occur at the peak of the curve. The 'mode' or most common measure is about 3 inches below the mean, a distribution common to most rainfall records. The reason is that there is a fixed lower limit (no rainfall), but no upper limit for the monthly falls; and therefore the mean is raised above the mode by an occasional excessively heavy fall.

What is the desirability of accepting the mean monthly rainfall as a standard for a particular region? The mean in the Taiping example is such that 70% months lie within the range 8 inches below the mean to 8 inches above it. The mode gives the same percentage of months lying within a range of only 7 inches above and 7 below. Thus to describe the rainfall for any month of the year within a 70% accuracy, the mode is better than the mean.

Although this principle has no great application when all the months of the year are considered together (as in the Taiping example), it may be of importance to agriculturists who wish to know the probability of the rainfall in one particular month falling

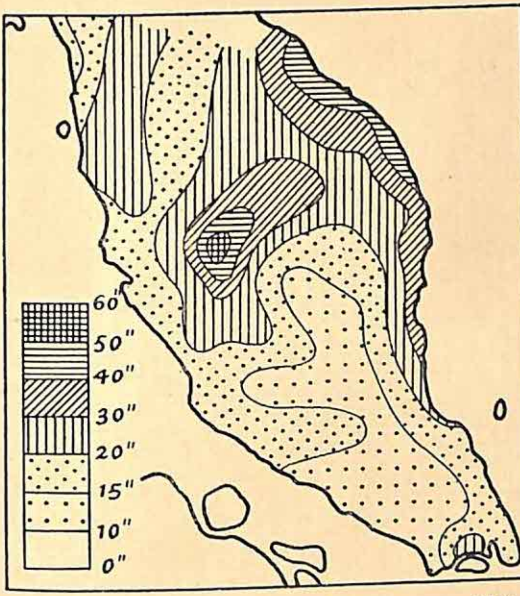


FIG. 31 Range of Monthly Rainfall, November—Malaya

within certain limits. If there are records over a long period (one hundred years or more), the probability may be calculated for each individual month of the year, preferably using the mode as the standard.

Visher[36] advocates considering range in assessing the value of mean rainfall figures, demonstrating with seasonal maps of the difference between the absolute maximum and absolute minimum rainfall in different regions. To benefit from this method, long-period observations are desirable. Its applications in the Malayan region are shown here (Figs. 30–32), but the results must be treated with a certain reserve, since most observing stations have been reporting for only twenty to forty years.

During April (Fig. 30) winds over Malaya are light and variable in direction, as the Northeast Monsoon has ended and the Southwest Monsoon not yet started. In such calm conditions, instability increases due to surface heating, and most rainfall of April consists of local showers. Fig. 30 shows that, except on parts of the east coast and over a strip of high country in the west, the range is small—between 10 and 15 inches. If we now refer to Fig. 25, we find that the mean April rainfall in the region is mostly 8 to 12 inches, and therefore we must assume that this mean is compounded

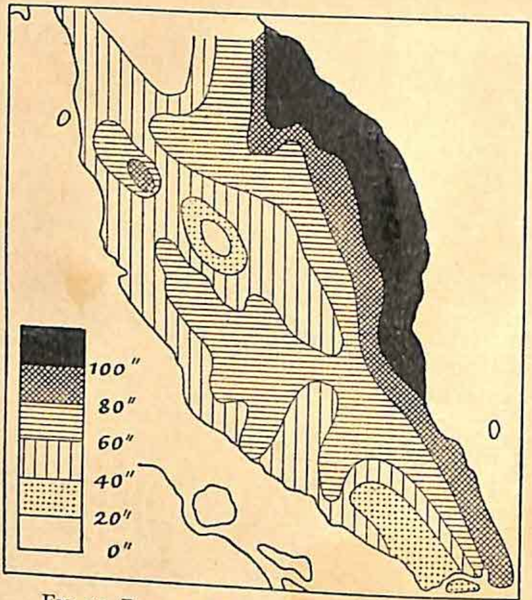


FIG. 32 Range of Annual Rainfall—Malaya

of monthly rainfalls varying mainly from about 4 to 16 inches $\left(\frac{8 + 12}{2} \pm \frac{10 + 15}{4}\right)$. Similarly, monthly totals in the western belt are mainly between 5 and 23 inches.

Ranges are greater in November (Fig. 31), particularly on the east coast and in the high central country. During November, the first rains of the Northeast Monsoon occur, and districts exposed to the northeast experience a change from showery conditions to those of general heavy rain. A delay in the onset of the monsoon may cause an abnormally low November rainfall, and an early monsoon commencement a high one.

Details of annual range are also shown (Fig. 32) to assist in the interpretation of the map of mean annual rainfall. As might be expected, there is a correlation between the magnitude of the annual rainfall and the magnitude of the range of annual rainfall totals.

The large rainfall range in the period of one month indicates that on occasions there may be very heavy falls of short duration. They may be of a few days' duration (the passage of the boundary dividing two air-streams), or perhaps a torrential thunderstorm over a few hours might expand the rainfall total to a value far above

the monthly mean. This is best investigated by considering the intensity or maximum falls which have occurred during a day.

4. Intensity of Rainfall

From 1931 to 1940 practically all the values of maximum rainfall in 24 hours at each of forty-seven stations in Malaya were between 2 and 8 inches. These are heavy falls, especially by comparison with those of most temperate countries. (The maximum 24-hour fall ever recorded at Ben Nevis was 7·75 inches.) A few stations had much heavier falls, as follows:

Pekan	19·18 inches	(Dec. 1938)
Penang Hill	14·29 ,,	(May 1939)
Kuala Krai	12·15 ,,	(Jan. 1939)
Port Dickson	12·72 ,,	(July 1935)
Kuantan	11·05 ,,	(Nov. 1931)

As the total rainfall at Pekan during December 1938 was 41·19 inches and the December mean is only 26·78 inches, this fall of only one day's duration had a great influence on the range of December total falls. In a similar way, the day's fall of 14·29 inches at Penang Hill considerably expanded its May range, since the mean monthly rainfall is only 14·95 inches. The precipitation at Pekan was probably due to an extensive disturbance in the Northeast Monsoon. Penang Hill's heavy fall was partly due to relief, though the leading boundary of the Southwest Monsoon may have been an influence because it crosses Penang during May. There is no such simple explanation of Port Dickson's heavy fall during July, when the Southwest Monsoon is always well established. It might derive from a large disturbance moving within the southwesterlies or, more probably, from an isolated thunderstorm.

Although some of Malaya's maximum 24-hour falls appear high, similar falls are not unknown in other parts of the region. Braak[37] reports that the highest daily fall ever recorded in the Indies was 20·12 inches at Besokor. Even this value is not high by comparison with the following 24-hour maxima [37, 38]:

Cherrapunji	40·79 inches
Funkiko (Formosa)	40·71 ,,
Honomu (Hawaii)	31·97 ,,

With 24-hour falls occasionally exceeding 19 inches in the Indies, there must have been even more intense falls of shorter duration.

The highest rainfalls for certain periods at any one of ten stations near Batavia from March 1923 to January 1925 were as follows[39]:

	Minutes		Hours								
Duration of Fall	10	30	1	2	3	4	5	10	12
Rainfall (inches)	0·99	2·32	2·92	3·50	4·49	5·40	6·22	7·40	7·44

That these figures are fairly representative of the surrounding region may be seen by a comparison with records in Ceylon,[40] and with the following observations of absolute maximum intensity at Singapore over the longer period from 1931 to 1941 inclusive:

Duration (hours)	¼	½	¾	1	2	3	4	5	6
Rainfall (inches)	1·56	2·58	3·93	4·73	5·79	5·84	5·84	5·84	6·23

Braak[41] has shown that the means of the maximum intensities of six selected stations in the Indies are not unduly higher than intensities recorded in Prussia.

5. Variation of Rainfall with Height

Rainfall is on the whole greater in high country than over plains. In any general survey of a region, including both hills and lowlands, much detail is necessarily neglected. Rainfall increases with height to a certain altitude and thereafter decreases. The rates of the increase and decrease vary considerably, as do the heights at which the change takes place.

In Java[42] rainfall increases from the coast inland over the plains, and also up to an altitude of 2000 to 3000 ft. Above these levels rainfalls may remain constant or show a small decrease, which above about 5000 ft. becomes a steady decrease. These rules apply to seasonal and annual totals and also to the mean 24-hour maxima at the various levels. On the other hand, the means of the 24-hour absolute maxima for a number of stations at each level remain constant for all heights. De Boer[42] suggests this arises because, although the number of rain-days and the duration of showers increases with height above 4000–5000 ft., the intensity of rain decreases, producing a compensation.

6. Diurnal Variation of Rainfall

Rainfall also has a varied distribution throughout each day. Maximum frequency of precipitation normally occurs during the afternoon over land and in the early morning over sea. These principles convey the impression that, at any tropical land-station, cloud increases during the morning to develop showers in the afternoon. This is true for most inland stations, but the early morning maximum over the sea has a pronounced effect on many coastal stations where at some seasons showers are far more frequent in the morning than in the afternoon.

Figs. 34 and 35 show the frequency of occurrence of precipitation at various times of the day in different parts of Malaya. The times of day are plotted against the percentages of days of the month on which rain falls during each hour.[43]* In such analyses a rain-hour is taken to be one on which there is a fall of 0·01 inch or more. Similar analyses adopting various values of minimum hourly precipitation have indicated that highest frequencies of precipitation occur at about the same time of day irrespective of the minimum standard adopted (Fig. 33).

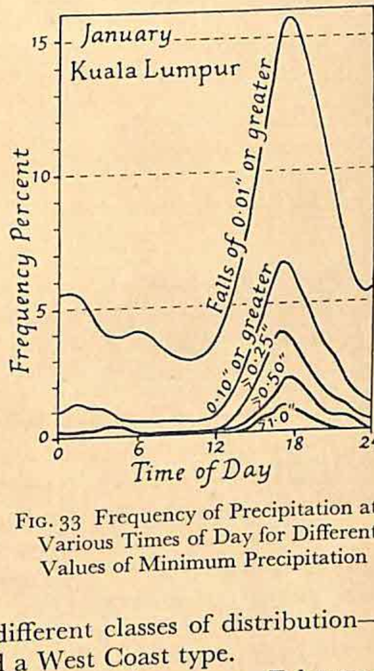


FIG. 33 Frequency of Precipitation at Various Times of Day for Different Values of Minimum Precipitation

The varieties of time of day of maximum frequency of precipitation over Malaya indicate three different classes of distribution—an East Coast type, an Inland and a West Coast type.

That at Kota Bharu is typical of the East Coast type. February, March and April show no marked peak in frequencies: fewest falls are experienced in the afternoon and most (particularly farther south) occur in the early morning. From May until October this timing is reversed and minimum frequencies are shortly before midday, while the maximum frequencies are in the afternoon, mainly about 6 p.m. (M.T.), but as early as 3 p.m. in the far south.

* These frequencies are calculated from fifteen years of autographic records, and the curves smoothed by plotting at each hourly interval the average of three consecutive hours.

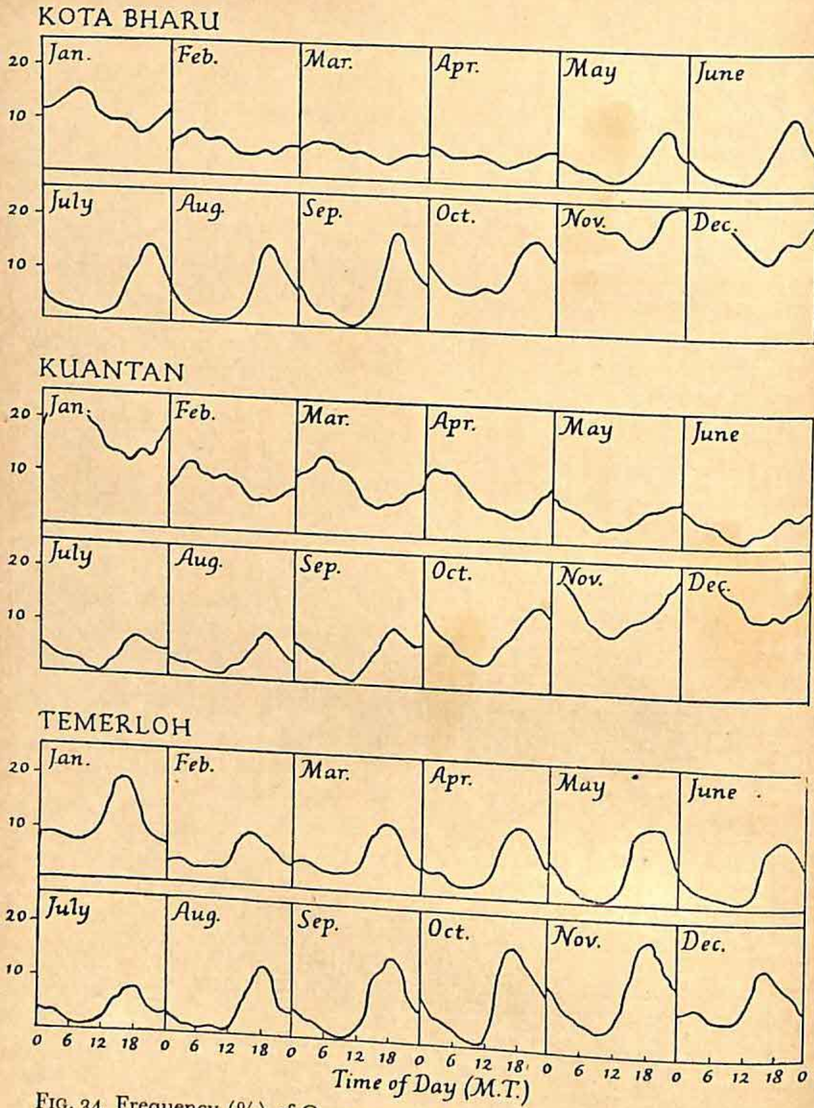


FIG. 34 Frequency (%) of Occurrence of Precipitation at Various Times of Day—East Coast and Inland Malaya (After Lea[43])

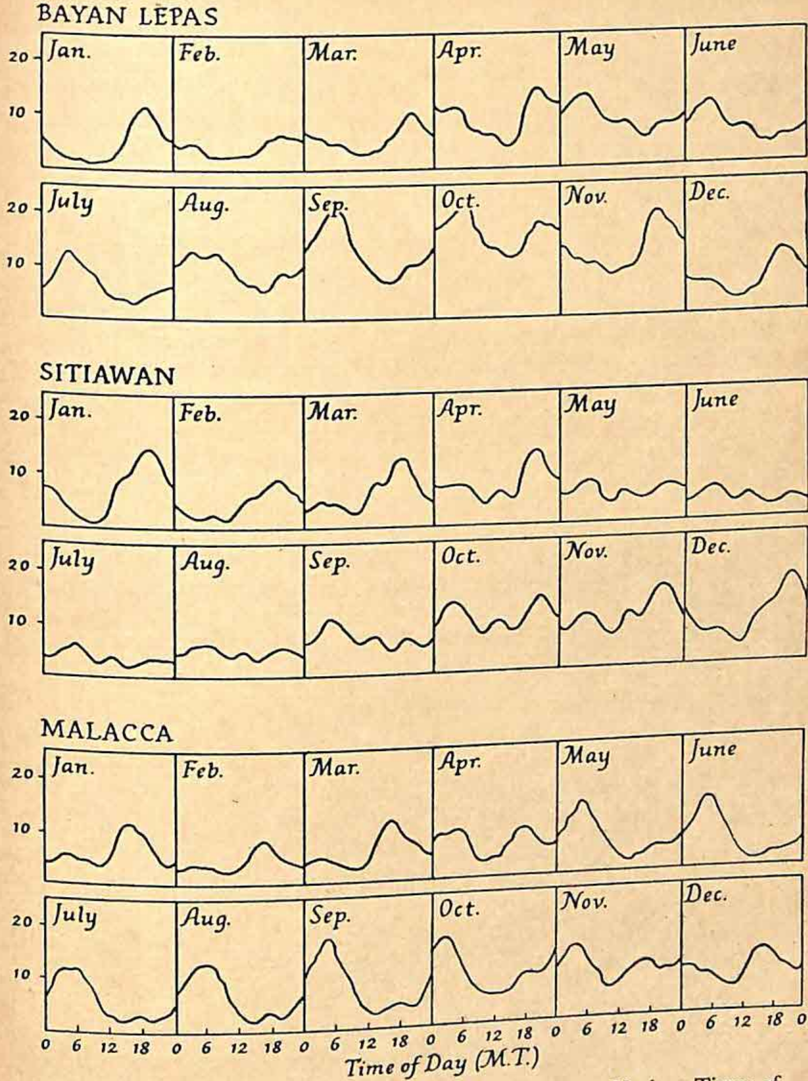


FIG. 35 Frequency (%) of Occurrence of Precipitation at Various Times of Day—West Coast of Malaya (After Lea[43])

The peak of maximum occurrence is later than midnight after October and 6 a.m. in December and January.

These changes are closely linked with seasonal upper-winds. During February, March and April (the 'inter-monsoon period'), light variable winds yield a fairly even rainfall distribution of showers from afternoon cumulonimbus cloud, together with showers from the sea in the early morning. After the start of the monsoon in May, winds are from the southwest and showers from the sea do not intrude. Precipitation then occurs in afternoon showers, but cloud development is delayed by a föhn effect in the southwesterlies (which cross the ranges), maximum frequency being in the late afternoon. The change to the Northeast Monsoon later in the year once more brings showers formed over the sea, their rain falling in the night or early morning as they drift to the coast and possibly develop orographic uplift.

The distribution inland is much simpler. Throughout the year, most stations show a period of negligible frequency of precipitation at about 10 a.m. The peak is always high, coming mostly about 4 p.m., but retarded until about 8 p.m. at some inland stations exposed to the Northeast Monsoon. The distribution at Kuala Lumpur is unique, with a small maximum from 5 a.m. to 6 a.m. from July to November, resembling the West Coast type.

The régime of the west coast is most pronounced at Malacca (Fig. 34). From December until March it has a simple diurnal cycle with maximum frequencies about 5 p.m., associated with light winds and diurnal heating of the land. Similar conditions apply over other western districts, although the peak is an hour or two earlier towards Singapore and an hour or two later in the north. The afternoon peak of frequencies at Malacca decreases slightly during April, and an equally prominent maximum occurs between 1 a.m. and 6 a.m. In May this early morning maximum is well developed at 6 a.m., with very little precipitation at other times of the day—conditions which continue for some months. A slight afternoon maximum is noticeable from August onwards, and this increases until, with the decrease of morning rains during December, afternoon showers again dominate.

Similar régimes may be found in varying degrees all along the west coast. That at Penang and Alor Star closely follows that at Malacca. Between Penang and Malacca, the morning maximum is less pronounced and afternoon showers well developed by October. The early morning maximum at Kuala Lumpur appears due to coastal showers moving inland.

The morning frequency of precipitation on the west coast occurs

when the Southwest Monsoon covers the country. Then, cumulus and cumulonimbus develop in the Straits of Malacca on most nights, and extensive cumulonimbus clouds commonly move to the west coast of Malaya in the early morning. The structure may be a single cumuliform column or a long line of cumulonimbus parallel to the coast. Sometimes its advance is accompanied by a strong squall termed a 'Sumatra' because it approaches Malaya from that direction. The formation of 'Sumatras' has such an effect on the weather of western Malaya and the Straits that it will be discussed fully in Chapter XIII. These squalls form over the Straits of Malacca at night, the radiational cooling of cloud tops probably helping to provide the instability for cumulonimbus formation. Furthermore, the outflow of cooled night air from the hills promotes the uplift of the unstable air over the straits by undercutting.[12]

There are many ways in which the rainfall of a region may be analysed. A detailed analysis may normally be made only for a single station or for a very small area, as such large variations are brought about by differences in topography and in the heat-absorbing properties of adjoining surfaces. Mean monthly and annual rainfall maps are only approximate guides to the actual conditions. Their reliability is a function of the number of reporting stations. Over large areas of Southeast Asia, particularly in the ranges of Malaya and Borneo, there are few records of rainfall, so that its regional rainfall maps are considerably less reliable than those of more closely settled regions of the world.

CHAPTER VII

Theory of Equatorial Air-movements

In middle latitudes a workable relation may be formulated between the spacing and orientation of the isobars and the speed and direction of the wind. This pressure–wind relation becomes less exact with decreasing latitude, until near the Equator it has no application at all. Hence developments in low-latitude meteorology have ignored the concept and have proceeded on different lines.

1. Pressure-Wind Relationship

If the earth did not rotate, the only force acting on any particle of air would be that due to the pressure gradient,* and consequently wind streams would be outflows across the isobars from higher to lower pressure. The effect would be to decrease the pressure differences. On a rotating earth, however, the streams are complicated by the force due to rotation.

Consider the case of a turn-table revolving counter-clockwise at angular velocity ω (Fig. 36). Suppose that a ball is thrown horizontally from the centre X of the table towards a distant point, P, outside the revolving table. The ball would appear to come directly towards an observer at point P; but to an observer at some point, A, moving with the table, the ball would appear to be deflected to the right. This deflection may be accounted for by the force which would produce an acceleration $2V\omega$, where V is the velocity of the ball. The direction of the acceleration would be perpendicular to the path of the ball in the plane of the turn-table. A stream of air near the North Pole would be deflected to the right like this, but at the South Pole the deflection would be to the left.

The deflective force, known as 'Coriolis Force', is significant at latitudes away from the Poles, and the acceleration due to it at latitude ϕ is $2V\omega \sin \phi$. At the Equator the value of this acceleration is zero.

Geostrophic Wind

If a particle of air, initially at rest, begins to move across the face

* Pressure gradient—difference of pressure per unit distance perpendicular to the isobars.

of the earth towards lower pressure, it is deflected more and more to the right (in the Northern Hemisphere) until the movement is perpendicular to the pressure gradient—i.e. along the isobars. Hence, systems of highs and lows persist with air rotating around their centres. The sense of the rotation in the Northern Hemisphere is clockwise around a high and counterclockwise about a low. These directions are reversed in the Southern Hemisphere.

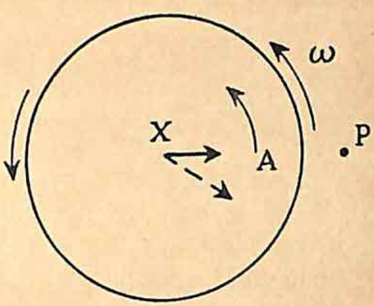


FIG. 36 Deflection of a Moving Particle

The acceleration due to pressure gradient is G/ρ, where G is the pressure gradient and ρ is air density. By equating the acceleration due to the earth's rotation with that due to the pressure gradient, a value may be found for the resultant wind speed along the isobars:

$$G/\rho = 2V\omega \sin \phi \quad . \quad . \quad . \quad . \quad (1)$$

$$\text{or} \qquad V = \frac{G}{2\rho\omega \sin \phi} \quad . \quad . \quad . \quad . \quad (2)$$

Underlying these relations between pressure and wind are assumptions[44] that:

(*a*) There is no acceleration or retardation of the system.
(*b*) The isobars are evenly spaced parallel straight lines which are not changing with regard to time.
(*c*) Friction may be neglected, the equation applying to an upper level where the frictional effect of the earth's surface is not effective.
(*d*) Movement is over a very small range of latitude so that $\sin \phi$ may be treated as constant.

The 'geostrophic wind' (V) from this equation approximates to the actual wind at middle and high latitudes. It applies particularly in areas free from local topographic effects and to levels from 1500 to 3000 ft., but not near the ground, where surface friction upsets the balance between the forces.

In temperate latitudes, meteorologists use scales based on equations (1) and (2) to measure, from the spacing of the isobars, the geostrophic wind velocity over any place, separate scales being computed for different scales of weather chart and for different

latitudes. Close isobars show steep pressure gradients and strong winds.

Consider now the geostrophic wind equation $V = \frac{G}{\rho f}$ (see (1)), where $f = 2\omega \sin \phi$ (known as the 'Coriolis Parameter').

Since ω is 2π radians in 24 hours or $7{\cdot}27 \times 10^{-5}$ radians per second, then $f = 2\omega \sin \phi = 1{\cdot}454 \times 10^{-4} \sin \phi$ and has the following values at the various latitudes.

ϕ (latitude)	0	10	20	30	40	50	60	70	80	90
$f \times 10^4$ sec.	0	0·252	0·497	0·727	0·935	1·114	1·259	1·366	1·432	1·454

The changes in f with latitude are evidently most rapid near the Equator and least towards the Poles.

Consider the case of a geostrophic wind of 25 m.p.h. (11·17 metres/sec.) at latitude 40° N. or S. Assuming that $\rho = 12{\cdot}93 \times 10^{-4}$ gm. cm.$^{-3}$, then the pressure-gradient equivalent to the geostrophic wind ($G = 2V\rho\omega \sin \phi$) will be 1 millibar per 50 miles. At low latitudes, however, very different gradients are found. For a geostrophic wind of 15 m.p.h., a common one at latitude 10°, the value of G is 1 millibar per 280 miles.

For decreasing latitude the Coriolis Parameter decreases rapidly so that a given value of the geostrophic wind is equivalent to a considerably greater isobaric spacing until, at the Equator where the parameter is zero, isobaric spacing is infinite and the equation no longer applies. Isobars are characteristically widely spaced near the Equator even when they are drawn at 1-millibar intervals; they are rarely straight or in the easy curves and simple patterns of temperate-latitude isobars. It is unusual to find two isobars even nearly parallel, and a representative gradient cannot be estimated with any accuracy. Hence, although the geostrophic wind equation is a most valuable one in temperate-latitude meteorology, it has little use in equatorial latitudes.

Gradient Wind

A further factor must be considered in pressure-wind relations. If the air is moving in a curved path, a centrifugal force will be acting radially, and must be compounded with the pressure-gradient force to balance the Coriolis force.

In an anticyclone, the pressure-gradient force is outward from the

centre in the same sense as the centrifugal force, and the equation is:

$$\frac{G}{\rho} + \frac{V^2}{r} = 2V\omega \sin\phi \quad . \quad . \quad . \quad (3)$$

where r is the radius of curvature of the path of a particle of air about the centre of the anticyclone. (Mass does not enter the equation, as only a unit of mass is considered.)

In a depression the pressure-gradient force is towards the centre of the depression and opposite to the centrifugal force. Therefore the equation is:

$$\frac{G}{\rho} - \frac{V^2}{r} = 2V\omega \sin\phi \quad . \quad . \quad . \quad (4)$$

These equations are quadratics and from them V can be determined:

$$V = \frac{r}{2}\left[f - \sqrt{f^2 - \frac{4G}{\rho r}}\right] \text{ for an anticyclone * } \quad . \quad (5)$$

$$V = \frac{r}{2}\left[\sqrt{f^2 + \frac{4G}{\rho r}} - f\right] \text{ for a depression } \quad . \quad . \quad (6)$$

Equations (3) and (4) are the geostrophic wind equation (1) modified by the inclusion of an expression for the centrifugal force which is called the 'Cyclostrophic Component'.

If, in equations (3) and (4), the value of r is taken as infinite, both equations reduce to the geostrophic equation (1), which is thus appropriate for curves of infinite radii (i.e. for straight isobars).

The value of V found from equations (5) and (6) is termed the 'Gradient Wind' and represents the value of wind speed along curved isobars.

The significance of equations (3) and (4) ($G/\rho \pm V^2/r = 2V\omega \sin\phi$) must be examined for low latitudes. If cyclonic curvature is great, the cyclostrophic term has importance even at latitude 10°,

* The solutions of the quadratic are $V = \frac{r}{2}\left[f \pm \sqrt{f^2 - \frac{4G}{\rho r}}\right]$.

The pressure gradient may be replaced by $V_{geost.}\,\rho f$ so that $V_{grad.} = \frac{r}{2}\left[f \pm \sqrt{f^2 - \frac{4 f V_{geost.}}{r}}\right]$.

Then, as the gradient wind should approximate to the geostrophic wind with increasing radius, the negative solution must be taken. The discarded root applies to systems of very small radii rotating in the opposite sense.

where the gradient wind corresponding to a geostrophic wind of 15 m.p.h. is:

12 m.p.h. for a radius of curvature of 500 miles.
11 m.p.h. for a radius of curvature of 300 miles.
8 m.p.h. for a radius of curvature of 100 miles.

The gradient wind equation applies to tropical cyclones even within the latitudes 10° N. and S. In them the Coriolis force may be significant because, although sin ϕ is small, V is appreciable. The ratio of the cyclostrophic term to the Coriolis term, however, is usually considerable (Fig. 37, after Byers[45]).

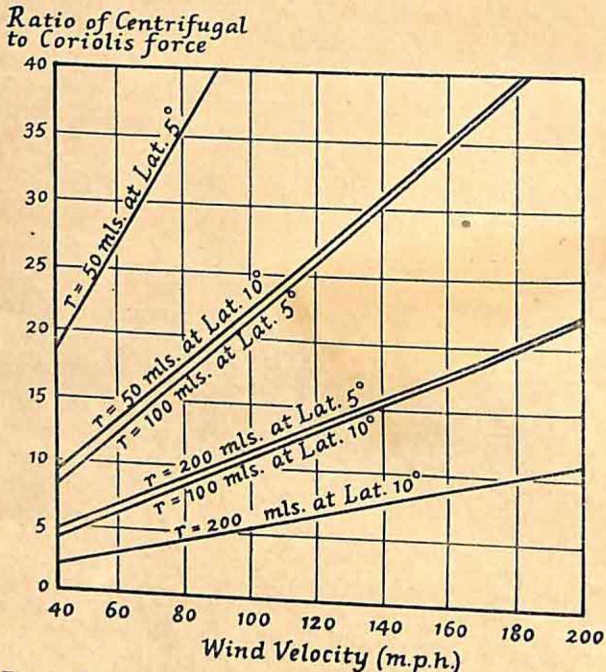


FIG. 37 Ratio of Cyclostrophic to Coriolis Term (After Byers)

Why is it that the gradient wind equation has, apart from a limited use in cyclones, no general application to the equatorial region? It appears that the simplifying assumptions (page 73) are not valid at low latitudes. We have seen that it is wrong to assume that sin ϕ is constant since it varies most rapidly for small values of ϕ, but let us study the effects of friction and accelerations in more detail.

Frictional Effects

Friction at the earth's surface acts in a direction opposite to the motion of the air, causing velocity to be diminished. The geostrophic term ($2V\omega \sin \phi$) is thus lessened, while the cyclostrophic term (V^2/r) is greatly decreased. Great friction may cause the pressure-gradient force to exceed the centrifugal force, and air to cross the isobars towards the lower pressure. Surface winds may deviate then greatly from the geostrophic-wind direction. The amount of deviation varies with the roughness of the ground and is greater at low latitudes than at high latitudes.

The following are mean values for the angle θ between the wind direction and the isobars: [44, 46]

Latitude (S)	5	10	15	20
θ	45°	41°	32°	31°

The effect of friction is greater over land than over the sea, and where the country is mountainous the deviation may extend to great heights.

The frictional effect is distinctive in cyclones, which usually fill when moving from sea to land. If it be assumed that the terms of the gradient-wind equation were in balance over the sea, the considerable reduction in the cyclostrophic term (V^2/r) over land will allow a large influx of low-level air across the isobars under the influence of the pressure gradient. Rising pressure at the centre of the cyclone ensues, followed by a reduction of pressure gradient.

Influence of Accelerations

The assumption that accelerations in the air-flow may be neglected must also be examined in applying the gradient-wind equation to low latitudes. In temperate and high latitudes, air-streams follow curved paths in systems which do not alter greatly over 12 or 24 hours. The accelerations of a parcel of air following the slow changes of gradient with altering pressure patterns are slight, and usually may be neglected there except for rapidly deepening or rapidly filling depressions. But at low latitudes, both gradients and winds are slight and, as it is difficult to measure the former, large alterations may occur without their significance being seen by the meteorologist.

Probably the most important factor causing changes of gradient

is the differential heating of land and water which can produce local variations of pressure up to 1 millibar.[47] Crossley points out that a reduction of pressure over land by only 0·5 millibar due to diurnal heating when the pressure gradient is 1 millibar in 300 miles, can shift the isobar 150 miles. Differential pressure changes of this small order have a great effect on winds which locally may become strong enough to dominate the flow-pattern at low levels. Over the Peninsula of Malaya diurnal heating frequently produces an area of low pressure during the early afternoon, causing winds up to about 3000 ft. to converge inland. Gradients and accelerations are also distinctly affected by the twice-daily variation of pressure, with an accompanying variation in the wind.

Low-latitude gradients continually change; the flow of air is never steady, so that the assumptions made in formulating the gradient-wind equation are not valid there. A new equation is needed for gradient wind in the equatorial region, but the problem is complex owing to the many variables involved, and no solution has yet been found simple enough to be practical. Specialised studies of the problem have been made, the most notable being that of Grimes,[48] which has been modified by Crossley.[47]

Movement of Air Across the Equator

Grimes takes as his assumptions for analysing air movement that the motion is a steady horizontal flow independent of longitude, and that the air involved has constant density without influence of frictional forces.

From the equations of motion, and taking specific initial values of flow, Grimes derives typical stream-lines for air-flow between latitudes 5° N. and 5° S. His solutions are by no means general, and Crossley[47] states that they apply to mean motions and not to actual motions. Forsdyke[44] has pointed out that by restricting the study to horizontal motion of air of constant density (i.e. to non-divergent and non-convergent motion), the solutions have little value because divergence and convergence are too important to be ignored.

Crossley[47] shows examples of stream-lines in relation to isobaric patterns (Fig. 38). They deviate greatly from the isobars. In the first case the wind direction is initially geostrophic, but, after crossing the Equator, the stream-lines make a large angle with the isobars. At a remoter latitude, the wind direction returns to the geostrophic. In the second case there is a trough at the Equator to which the isobars are parallel. The air moves in a straight path across the trough, and then curves round into the higher pressure. Thus air flowing from one hemisphere to the other may pass completely

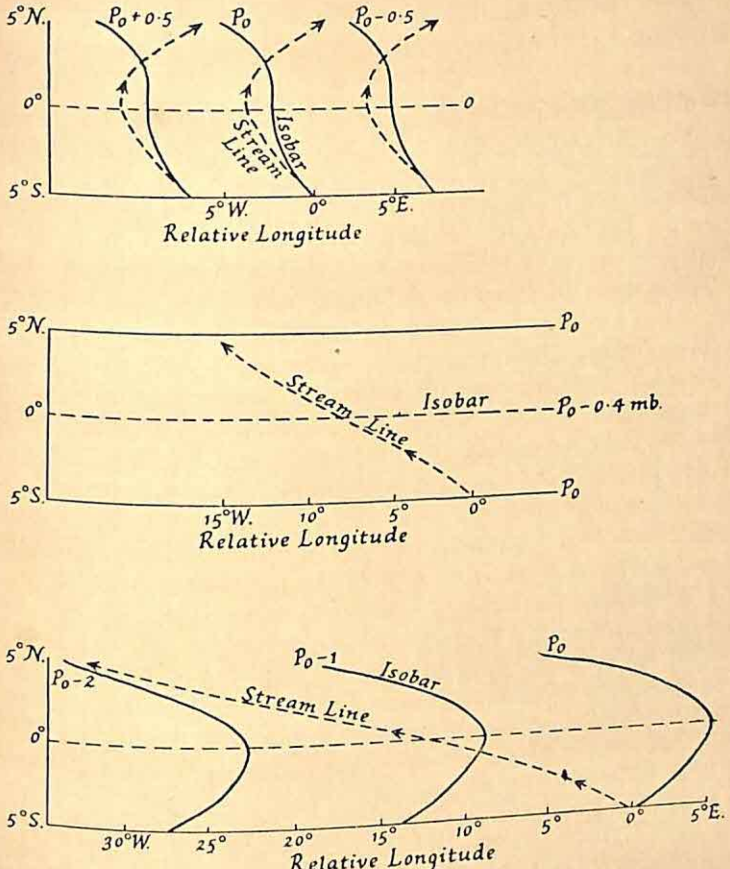


FIG. 38 Stream-lines and Isobars (After Crossley)
In each case the wind at Lat. 5° S. is S.E. 600 cm./sec.

across the equatorial low-pressure trough. In his third case there is also a trough along the Equator, with winds which, initially at a wide angle to the isobars, rapidly become geostrophic.

2. Isallobars

When a pressure distribution is moving or altering, the changes may be detected by drawing 'isallobars'—lines of equal pressure-change over a period (Fig. 39). Isallobaric patterns resemble isobaric patterns, forming in closed curves around areas of greatest rise or fall of pressure. When an isallobaric low occurs near an isobaric depression, the movement and possible deepening of the latter may be determined from the intensity and relative position

of the isallobaric low. The changes and movements of highs may be predicted similarly.

At low latitudes, pressures are greatly influenced by:

(1) The semi-diurnal pressure variation.
(2) A diurnal variation arising from insolation contrasts (which are greater over land than over sea).
(3) Reducing pressure at a high-level station to its sea-level equivalent.

Hence, isallobars are drawn only for periods of 24 hours beginning soon after dawn. They rarely show movements of the shallow equatorial highs and lows, but they indicate intensification or collapse of these systems.

The movements of cyclones may be calculated from the isallobars by using a displacement formula.[49, 50]* As this method is one of extrapolation, allowance must be made for any acceleration or retardation of the cyclones, decided from an inspection of the pressure-tendency profiles. These profiles are constructed by plotting pressure changes at all places along a line passing through the centre of the cyclone, and preferably oriented in the direction of travel of the cyclone. Typical profiles are shown in Fig. 40 (after Byers[49] †). Tendency is plotted as ordinate (OY′) and distance from the centre of the pressure system is plotted as abscissa (OX′).

* Two axes are chosen intersecting near the centre of the cyclone, and the unit of length of these axes is chosen to be as great as possible but contained in the cyclonic system.

The component of motion along each of the axes may be expressed as:

$$S' = \frac{1}{2}\frac{T^{(1)} - T^{(-1)}}{(p^{(0)} - p^{(1)}) - (p^{(0)} - p^{(-1)})} \quad . \quad . \quad . \quad . \quad (7)$$

where $T^{(1)}$ and $p^{(1)}$ are the 24-hour pressure tendency and the barometric pressure at a fixed distance along the axis ahead of the pressure centre;
$T^{(-1)}$ and $p^{(-1)}$ are the values of these elements at a point along the axis at the same distance behind the centre;
$p^{(0)}$ is the pressure at the centre of the system;
S' is the distance travelled by the centre along the axis in a period of twenty-four hours, expressed as a fraction of the chosen length of the axis.

The components of motion are found at points along each of the intersecting axes, and perpendiculars are drawn from each of these points. Their intersection gives the position which, in the absence of accelerations, will be occupied by the centre of the system in twenty-four hours' time.

† Displacement Formula and Tendency Profiles by permission from *Synoptic and Aeronautical Meteorology*, by H. R. Byers, copyright 1937; McGraw-Hill, New York.

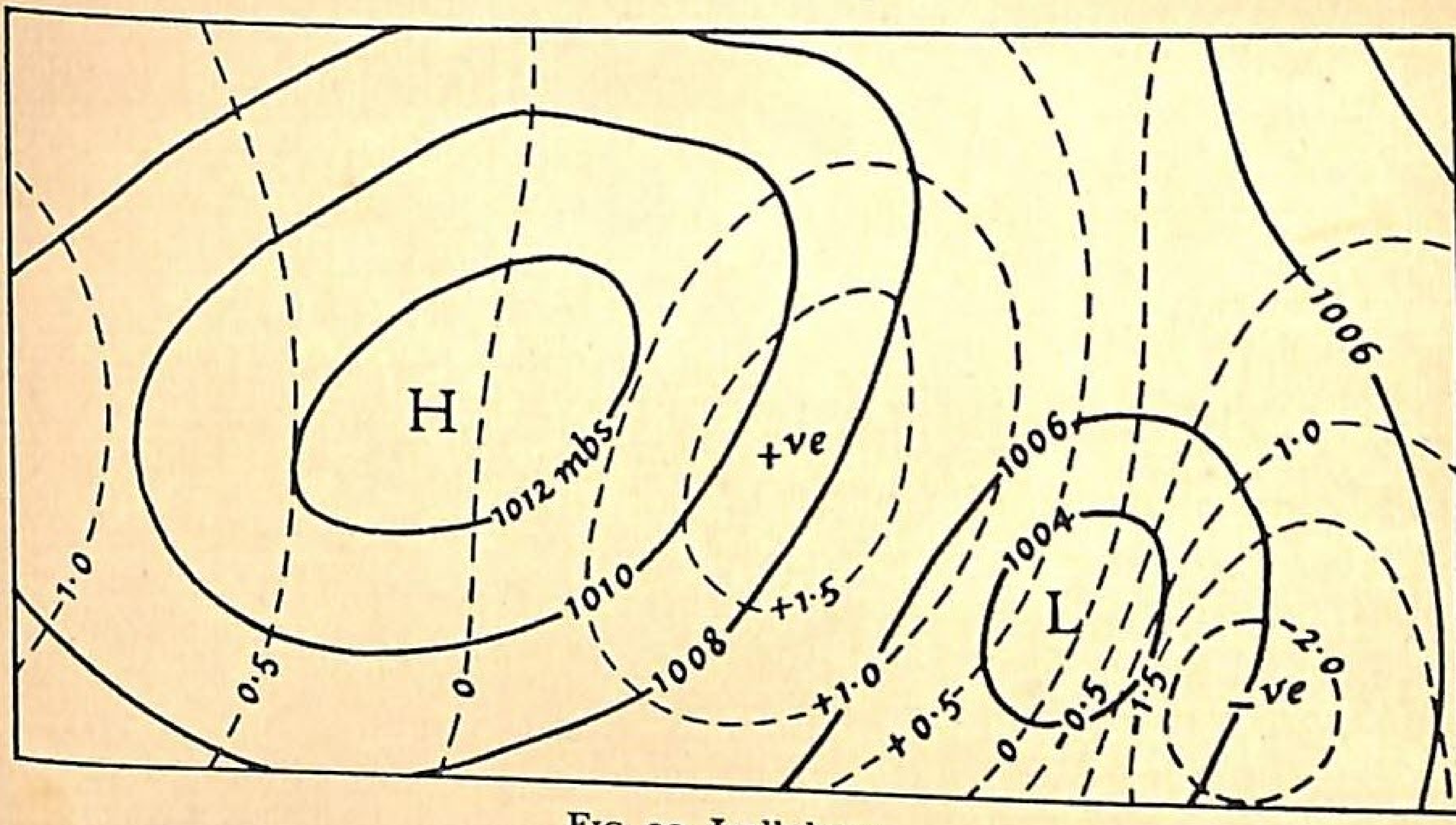


FIG. 39 Isallobars

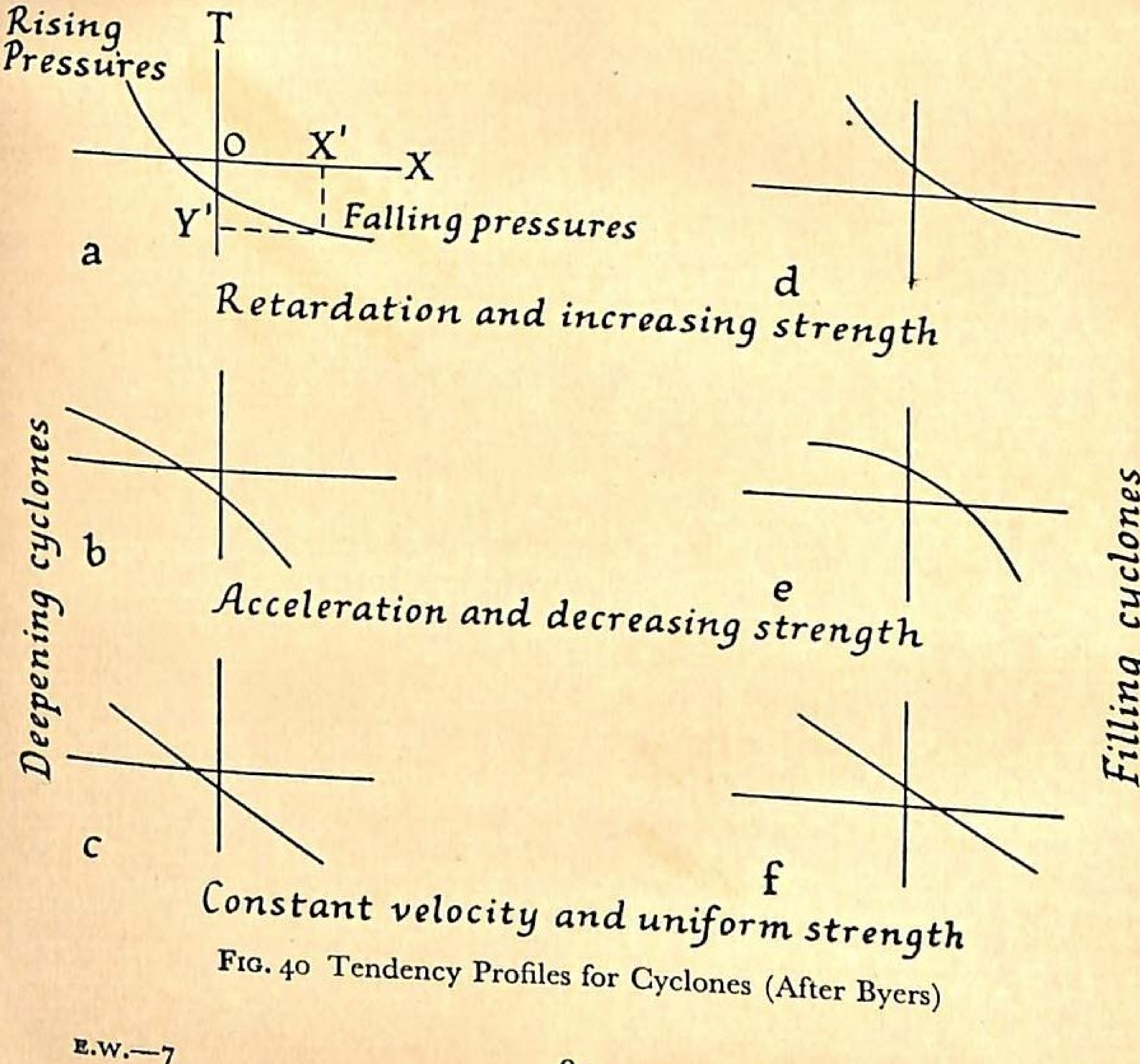


FIG. 40 Tendency Profiles for Cyclones (After Byers)

3. Winds at Higher Levels

If there are two distinct regions of equal pressure in the atmosphere, the ratio of the fall of pressure with height in one region to that in the other region varies inversely to the ratio of their temperatures.* Where temperatures are high, pressure falls off slowly, and where temperatures are low, the pressure-change with height is rapid.

Consider a cross section of the atmosphere along a meridian about middle latitudes of the Northern Hemisphere (Fig. 41).

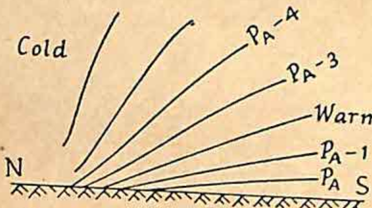


FIG. 41 Maintenance of Westerlies at Upper Levels

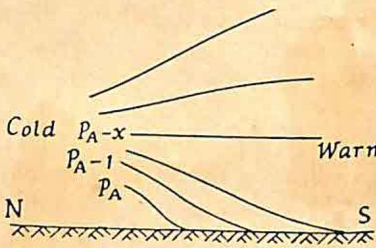


FIG. 42 Westerlies Above Easterlies

Isobars P_A, $P_A - 1$ (millibars), $P_A - 2$, etc., denote the pressure at various levels, and as surface pressure in this example is lower to the north than to the south, the wind, under geostrophic control, will be a westerly (blowing away from the reader). The air to the equatorial side will generally be warmer than that on the polar side, so that pressure falls off quickly with height poleward but more slowly equatorward. Thus pressure at upper levels is still higher on the equatorial side and the north–south gradient of pressure is intensified. Winds at upper levels are still westerlies, and (under geostrophic control) they increase with height in proportion to the increasing pressure gradient, amply illustrated in the 'Roaring

* $$\frac{\Delta p}{\Delta z} = -\frac{Pmg}{RT_m} \text{ in each region} \quad . \quad . \quad . \quad . \quad (8)$$

where $\frac{\Delta p}{\Delta z}$ is the change in pressure with altitude;

P is the pressure;
m is the gram-molecular weight of air;
g is the gravitational constant;
R is the universal gas constant;
T_m is the mean temperature of the air.

Forties', where westerlies at the surface invariably increase their speed and maintain their direction at upper levels.

Take the case along a meridian in the Northern Hemisphere, when surface pressure is lower to the south than to the north (Fig. 42). In this case the surface wind will be blowing easterly (towards the reader). As the air is warmer on the equatorial side than on the polar side, pressure will fall off more quickly with height in the north than in the south, a fall which will continue until, at some upper level ($(P_A - x)$ in the example), pressure between the two regions will be equalised. Up to this level of zero pressure gradient the easterlies must slowly decrease without necessarily changing direction. Farther above the level of equalised pressure, pressure decreases still more to the north than to the south, so that eventually a new gradient of pressure is built up with lowest pressures to the north. Winds at these upper levels, under geostrophic control, will be westerlies, and their strength will increase with height corresponding to the increasing horizontal pressure gradient.

Conditions in the Southern Hemisphere may be investigated similarly. In both hemispheres westerly winds increase with height up to the base of the stratosphere, and easterlies decrease with height, to be replaced at higher levels by increasing westerlies. There are exceptions in either hemisphere, when, for short periods and over limited areas, moving masses of air may upset the normal temperature distribution so that the temperature gradient may be oriented in a direction other than north–south, and the upper winds come not from the west but perhaps from northwesterly or southwesterly directions.

This shows that to determine the wind at upper levels another component must be applied to the geostrophic wind. This component (the 'Thermal Component'*), depends on temperature-distribution. Its direction is along the isotherms with low temperature on the left in the Northern Hemisphere, and on the right in the Southern Hemisphere.

To examine the upper flow at low latitudes, a cross-section of the atmosphere with north–south orientation across the Equator is

* The thermal component in any layer of the atmosphere may be calculated from pressure and temperature data, or obtained from a pilot-balloon ascent by plotting vectors for the wind of each level. For example, in Fig. 43 vectors are drawn from the common origin, O, to represent the wind in direction and speed at various levels. Then, the difference of velocity between any two levels, say between 3000 and 10,000 ft., is represented by the shear vector AB. This shear is due to the thermal component, being in the direction of the isotherms of mean temperature within the layer and of length proportional to the horizontal gradient of mean temperature through the layer.

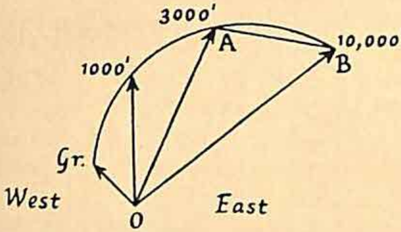


FIG. 43 The Thermal Component

Surface 140° 5 m.p.h.; 1,000 ft. 180° 10 m.p.h.; 3,000 ft. 210° 15 m.p.h.; 10,000 ft. 230° 20 m.p.h.

shown (Fig. 44). Mean pressures are high both north and south of the equatorial belt, illustrated by the concavity of isobar P_A. Mean temperatures are lower both to the north and south, so that the rate of fall of pressure with height is least at the Equator. Therefore, at some upper level ($(P_A - x)$ in Fig. 44), a zero horizontal pressure gradient is met. Mean low-level winds near the Equator are from the northeast in the Northern Hemisphere, and from the southeast in the Southern Hemisphere; at the altitude equivalent to $(P_A - x)$, the pressure gradient becomes zero and convergence of northeasterlies and southeasterlies ceases.

At high levels the horizontal pressure gradient is reversed so that pressure is highest over the Equator. Theoretically, at the equinoxes the upper air should first diverge in due northerly and southerly directions from the Equator. Then it should gradually assume a more westerly track with the increase in Coriolis force with increasing latitude. Eventually flow should turn to southwesterly in the Northern Hemisphere and to northwesterly in the Southern Hemisphere. This does occur, but the equatorial upper flow is by no means simple because the following factors exert a large influence:

(1) The surface pressure trough follows the ecliptic (path of the sun), so that latitudinal location of the low-level convergence and upper-air divergence varies with the seasons. Idealised patterns of the seasonal horizontal flow at the Equator in Fig. 45 are similar to the air-stream patterns at the surface and at 30,000 ft., shown in Chapters I and XI.

(2) Monsoonal effects may at times create a sufficiently strong flow to control wind directions to high altitudes.

(3) The horizontal north–south pressure gradient at low levels may be so great that the resultant inflow is strong, causing the northeasterlies and southeasterlies of the surface to persist to the base of the stratosphere.

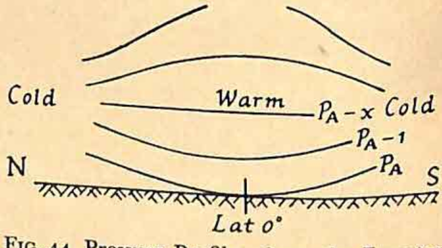


FIG. 44 Pressure Profiles above the Equator

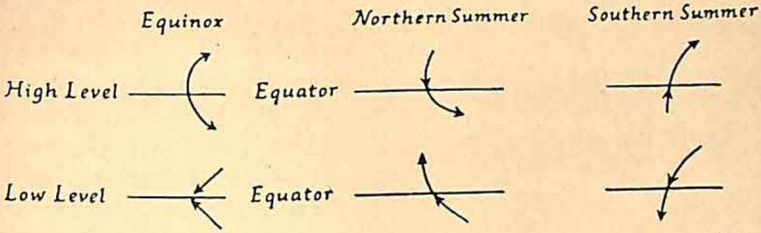


FIG. 45 Movement of Air across the Equator

Thus, though it is difficult to lay down rules for the upper flow at low latitudes, it is possible to interpret the day-by-day changes of flow in terms of the general principle modified by these influences.

Upper-air Charts

Upper-air charts are gaining popularity with meteorologists of temperate latitudes. Using them is based on the conception that major disturbances in the westerlies of the lower atmosphere are related to disturbances at upper levels of about 10,000 ft. The charts show isobars and isotherms drawn for the 10,000-ft. level (and for others) from radio-sonde and aircraft reports. Alternatively, the charts are made at particular isobaric levels (such as 750 and 500 millibars) and on them are traced lines joining places of equal height. The two methods are similar, and the same arguments apply to both.

A study of the 10,000-ft. chart for the temperate-latitude westerlies discloses that the isobars form into systems of waves involving troughs and ridges, which are either stationary or moving down the stream (Fig. 46).

For a stream running from west to east and containing a wave perturbation moving latitudinally, the velocity of the wave may be expressed[51, 52] by the formula:

$$C = U - \frac{\beta L^2}{4\pi^2} \qquad (9)$$

where U is the horizontal velocity of the particle in the undisturbed state;

L is the wave-length;

β is the rate of change of the Coriolis Parameter with increasing latitude northward.

From this equation it may be seen that:

(1) If C is positive $\left(U > \frac{\beta L^2}{4\pi^2}\right)$, the wave will move to the east.

(2) If C is zero, the wave remains stationary with air flowing through it along the isobars.

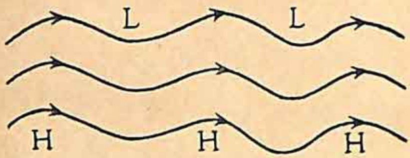


FIG. 46 Wave-disturbance in the Westerlies

If the stream in an undisturbed state is flowing latitudinally, the isotherms also lie along the parallels. On the introduction of a wave disturbance into the stream, any changes of temperature at a point may be considered due (if vertical motions are ignored) to the motion of the air particles along the isotherms. Under those circumstances it can be demonstrated mathematically[51, 53] that the relation between the amplitude of the isobars A_P and the amplitude of the isotherms A_T is:

$$\frac{A_T}{A_P} = \frac{U}{U - C} \quad . \quad . \quad . \quad . \quad (10)$$

where U and C have the same meaning as before.

From this equation it may be seen that the ratio A_T/A_P may be either positive or negative, and the negative values denote that the deformation of the isotherms is 180° out of phase with the deformation of the isobars.

Various cases which may arise can be investigated in the manner of Starr.[51]

(1) If $C = 0$, isobars and isotherms coincide (Fig. 47*a*). In other words, if the wave is stationary, the isotherms lie along the stream-lines.

(2) When C is positive but less than U, the ratio A_T/A_P is positive and greater than unity. Then the amplitude of the isotherms is greater than the amplitude of the isobars, but they are in the same phase. This is the case of a wave moving with moderate velocity (Fig. 47*b*).

(3) When C is positive but greater than U, the ratio is negative and the amplitude of the isotherms is greater than that of the isobars. Thus in a fast-moving wave troughs are warm and ridges are comparatively cold (Fig. 47*c*).

(4) If C is small and negative, the ratio A_T/A_P is less than unity. This is the case of a wave moving in a direction opposite to that of the stream, and then isotherms have less amplitude than the isobars, though both are in the same phase (Fig. 47*d*).

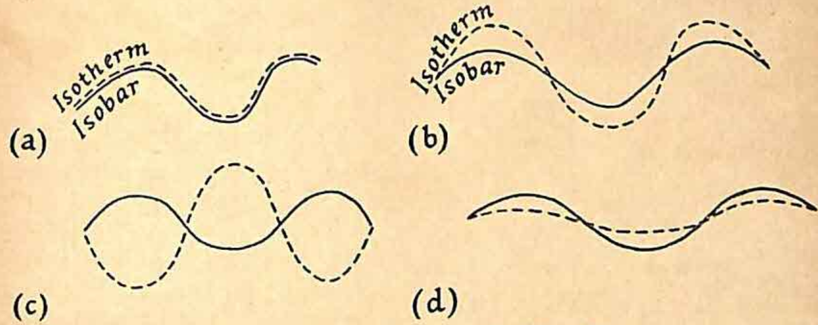


FIG. 47 Types of Waves in the Westerlies

(*a*) Stationary wave
(*b*) Wave moving with moderate velocity
(*c*) Fast-moving wave
(*d*) Wave moving slowly in direction contrary to stream-line flow

The theories outlined here also apply to the broad easterly streams of the tropical and equatorial regions. Riehl[54] has enlarged on the principles and produced examples of 'easterly waves' moving westward in the Caribbean Sea and the Gulf of Mexico. Further examples are afforded by Vuorela,[55] who describes seven such disturbances moving westward over the tropical Atlantic Ocean during a period of nine weeks. Riehl investigated the location of convergence and divergence with respect to easterly waves by quoting Rossby's[56] Vorticity Equation as follows:

$$\frac{f+\zeta}{D} = \text{Constant} \quad . \quad . \quad . \quad . \quad (11)$$

where f is the Coriolis Parameter, ζ is the relative vorticity (or spin) of a system of air particles with respect to the earth, considered positive for cyclonic vorticity, and D is the pressure difference between the top and the bottom of the layer.

Suppose that a particle begins to move from A to B along the stream-lines of Fig. 48 (Northern Hemisphere). As the particle is moving northward initially, f must increase. If D were to remain constant, ζ would have to become smaller and smaller and eventually negative in order to maintain the balance specified by $\dfrac{f+\zeta}{D} =$ Constant.

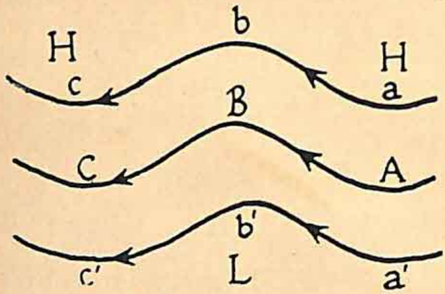


FIG. 48 Wave in the Easterlies

This means that the particle must develop anticyclonic curvature, but Fig. 48 shows that the particle actually develops cyclonic curvature. Therefore the premise that D remains constant cannot be correct and, if both f and ζ are increasing, D must also increase and there is convergence in the region between aa' and bb'. To the west of bb', on the other hand, f is decreasing and ζ is decreasing as its sign changes from positive to negative (from cyclonic to anticyclonic curvature). Hence D also must decrease and there is divergence between bb' and cc'.

The sequence of weather may be deduced from these theoretical considerations—fair conditions precede the trough, increasing cloud and perhaps precipitation follow the passing of the crest. Given a sufficiently good network of pilot-balloon stations for correctly drawing stream-lines, and given sufficient radio-sonde coverage to permit the drawing of isotherms, it should be possible to gauge the speed of the wave. Having estimated it and knowing the weather usually associated with a wave, weather changes may be predicted for places in its path. Owing to lack of observational material, it is not possible to apply these principles in the region of Southeast Asia. The surges of the Indian Southwest Monsoon and of the Northeast Monsoon in the China Sea are similar to easterly waves, but no record is known of a persistent travelling wave there.

CHAPTER VIII

Air-stream Analysis

The identification of currents of air from different sources by comparison of their temperature and moisture structures is rarely possible at low latitudes because all streams have such similar properties. Therefore a stream-line analysis is employed by which identification is made from reports of wind at various levels.

1. Air-mass Analysis

An air mass is a body of air with almost homogeneous properties covering an area of some thousands of square miles. The properties may either be similar throughout the mass or vary only gradually across it.

The temperature and humidity of the air mass are established by (1) a source region, and (2) transformations during travel.

The term 'source' is a relative conception, because air is in constant motion either horizontally or vertically, and nearly all regions could at some time be considered 'sources'. In general, polar regions and cold continents in winter are 'cold sources', and the tropics and warm continents in summer are 'warm sources'. Siberia may be considered as an air-mass source, because the outflow from an anticyclone centred there constitutes the winter monsoon of China.

The transformation of an air mass varies with the region being traversed and the time of travel. A rapid flow from polar to temperate regions undergoes less change than a slow-moving one, but far greater change than do the Trade-winds which, though slow-moving, flow latitudinally across regions of similar temperature.

Air masses are classified according to their sources or indirectly according to latitude and temperature; they may be called either 'cold' and 'warm', or 'polar' and 'tropical' masses. Differentiation of air masses takes place at the boundary between two masses, and classification then depends on comparing them.

Its source also affects the moisture content of a mass, and is usually described by the secondary divisions 'Maritime' and 'Continental' as listed overleaf:

Air Masses	*Symbol on Weather Map*
Tropical Maritime . . .	T_m
Tropical Continental . .	T_c
Polar Maritime . . .	P_m
Polar Continental . . .	P_c

In general, Tropical Maritime air is warm and very moist, while Tropical Continental air is warm and dry. Polar Continental air is cold and dry, as compared with Polar Maritime air which is cold and fairly moist. Most of the air masses in equatorial South-east Asia are Tropical Maritime (page 91).

Transformations of an air mass during its travel may be considerable. The suffixes 'Maritime' and 'Continental' may denote the modifications undergone after air has left its source, or the prefix N may denote a modified current (e.g. Modified Tropical Maritime Air—NT_m). The most important changes occurring in a mass are due to ground temperature in the region being traversed. Moisture content, at least in the lowest layers, is changed according as the surface over which it travels is land or sea. If the air is travelling over a warmer surface, the air temperature is increased in the lowest layer, whose capacity to hold moisture is thereby increased. Whether or not the moisture actually increases depends on the available supply.

The warming of air in its lowest layer during travel to warmer latitudes has other important effects, steepening lapse-rates and setting up instability. Air masses which enter the tropics attain lapse-rates exceeding the saturated adiabatic, and in many cases approaching the dry adiabatic. Lapses in excess of the latter, however, do not occur on a large scale because the high degree of instability tends to adjust the structure to a smaller lapse-rate through the interchange of air between layers by vertical currents.

Cooling in the lower layers of masses moving from tropical regions decreases their capacity to contain moisture and decreases or inverts the lapse-rate. Thus air of tropical origin entering temperate regions is notably stable. Surface cooling makes relative humidities high in the lowest layer, but as vertical currents are restricted by stability, the moisture distribution aloft is a function of the source region rather than the region being traversed.

In temperate regions, identifying air masses by comparing their properties of surface temperature, dew-point and lapse rate can be done with accuracy. This is not practicable in the equatorial region owing to the similarity of the masses.

2. Air Streams

The air masses which enter the equatorial portion of Southeast Asia have varied sources in both hemispheres. They are mainly maritime, being drawn from the western Pacific and from the Indian Ocean, but two continents also supply air to the region—Northeast Asia and Australia. A further source of less importance is Borneo,[57] where small, shallow anticyclones may become stationary long enough to be separate entities. In general, air may flow into Southeast Asia from any direction other than the north, and each direction is linked to seasonal changes. Currents from the north rarely enter the region because, although an extensive source exists in the Siberian High of the northern winter, the mountainous country of the Himalayas, of Burma and northern Thailand hinders a direct outflow of this air southward.

Air from the northeast may have a direct track through the China Sea or, as in the Northeast Trades from the Northwest Pacific, have a long passage across the waters surrounding the Philippines and the South China Sea. In both, the passage is across seas where temperature varies little other than the gradual latitudinal increase, and air temperatures in the lower layers tend towards those of the sea surface. Water evaporated from the surface while travelling over the sea increases the moisture content of the lower air, and dew points slowly rise.

Currents of air from the east to southeast—the Southeast Trade Winds of the Southwest Pacific—are maritime by origin and travel. Their route is nearly latitudinal across a wide expanse of tropical ocean: they become nearly saturated and, when entering the equatorial region, temperatures in their lowest layers are close to those of the sea.

Air from directions south of southeast sometimes approaches with a direct track from the Australian continent, as an outflow from a winter anticyclone centred there. Thus, to a slight extent, Australia may be considered a source region of air for Southeast Asia. But, even during winter, there is no permanent anticyclone over Australia persisting as long as those of Siberia. Palmer[58] has shown that the Australian region is crossed regularly by disturbances and migratory anticyclones. Moving from the west, these are composed of modified polar maritime air of Southern Indian Ocean origin, and hence are fairly moist initially. The air outflowing northward from an anticyclone temporarily covering Australia will have a humidity lowered by heating over Northern Australia, and perhaps by deposition in orographic rainfall on the

Victorian and New South Wales coasts. At low levels, this southeasterly stream may be warm and reasonably dry on entering the Indies: yet, to reach Equatorial Southeast Asia, it must cross one thousand miles of tropical sea, evaporation from which raises the dew-point almost to the air temperature, which in turn is governed by sea temperature. Therefore, despite its origin as a comparatively dry current from Northern Australia, it reaches Equatorial Southeast Asia as a warm moist current.

Air from the south does not enter the Malayan area. During the northern winter it is precluded by the Northeast Monsoon which blows farther south than Java. During the northern summer, conditions appear to favour air crossing the Equator from the south, but this is prevented by the breadth of the Indian Southwest Monsoon and by the persistence of the Southwest Pacific Trades across the Indies. Air travelling northwards along the West Australian coast is diverted counter-clockwise about the anticyclone lying to the west. Later, far out in the western Indian Ocean, it veers west on crossing the Equator[18] to join the Southwest Monsoon, losing all polar properties by the prolonged increase of heat and moisture.

Streams to Southeast Asia from the northwest to southwest, including the Southwest Monsoon, all have a long sea-track over the Indian Ocean, which makes them warm and moist.

During the northern winter, an anticyclone of varying intensity centres over India. A northerly outflow of warm, fairly dry air from its northeastern sector crosses the northern Indian Ocean: this air turns clockwise to northeasterly by latitude 10° N., crossing Ceylon as part of the anticyclonic circulation. A temporary intensification of the anticyclone may extend the flow as north-northwesterlies to reach northern Malaya, though its track over the sea has considerable effect in raising dew-points and temperatures. Intrusions of these Indian northwesterlies rarely pass south of 5° N. over the Malayan region because, at this season, the Northeast Monsoon is strong and unbroken.

In general, air streams (T_m) entering Equatorial Southeast Asia are warm and moist in the lower layers, although three may contain less moisture than others. These three are (1) southeasterlies outflowing from a migratory anticyclone over Australia; (2) Northeast Monsoon air of short track; and (3) Indian north-northwesterlies from the winter high.

Attempts by equatorial meteorologists to identify these air masses by surface properties have been disappointing. The three drier streams are not sufficiently dry to distinguish them and, although separate origins are clear from the orientations of the wind streams,

their surface temperatures and dew-points are similar, and diurnal and local variations in them are so great that they mask any slight differences.

Identification may be sought in the upper air where free from surface effects, as an air mass may be expected to retain aloft properties related to the source region. One serious disadvantage arises: all the masses have entered the region from cooler to warmer latitudes, causing their structure to tend to instability, absolute or conditional, and preventing identification by comparing lapse-rates.

The tendency to instability raises another difficulty. Suppose the moist air is conditionally unstable; a particle being disturbed upwards will cool at the saturated adiabatic rate. Because the atmosphere is unstable for rising moist air, there is at no level a natural stay to convection. Consequently, the high moisture content spreads to great heights, and even masses of continental origin will receive increases in their moisture content aloft, which tend to prevent identification of them as separate masses.

John[8] has analysed a series of upper-air observations at Singapore from December 1946 to July 1948 in an endeavour to identify air masses by their structure. The problem may be seen by considering the mean monthly tephigrams for January and July in Chapter III. Each of these represents a different stream. In January the Northeast Monsoon is well established, and there are no great fluctuations in its direction. In July the Southwest Monsoon is established and its direction fairly constant. Comparison of the two tephigrams will show that, despite their different origin, no very marked dissimilarities are evident in the upper-air structure. Changes of air stream from day to day usually consist of very slight alterations in the direction of flow, so that the likelihood of identifying them in the temperature-moisture structure is small.

John decided that air streams are not identifiable from lapse-rates, and he described the properties of all the streams affecting Singapore in the following general terms:

(1) Up to the altitude of 500 millibars, the lapse-rates are less than the dry adiabatic but greater than the saturated lapse-rate.

(2) In the lowest layers, the lapse-rates are small, being generally less than the saturated adiabatic lapse-rate.

(3) Inversions seem to be entirely absent, but there is a tendency for an isothermal layer from 1000 to 950 millibars in March and April, when nocturnal radiation is effective.

He classifies the streams affecting Equatorial Southeast Asia from a consideration of their trajectories (Fig. 49). The classification

is similar to that described earlier in this chapter, but varies in specifying the source by a suffix; e.g. Modified Tropical Australian —NT_A. The classification is as follows:

Latitude	*Nature*	*Symbol*	*Source and Modification*	*Remarks*
Polar	Continental	P_s	Northeast and Central Siberia	Reaches Malaya only after modification
		$NP_{s\ \{land\}}$	Modified over China and Indo-China	Seldom reaches Malaya without further modification over the sea
		$NP_{s\ \{cold\ sea\}}$	Modified over Sea of Japan, Yellow Sea, South China Sea	Reaches Malaya as a burst of the Northeast Monsoon in late December and in January
		$NP_{s\ \{warm\ sea\}}$	Modified over West Pacific and South China Seas	Reaches Malaya with properties similar to T_{NP} as a component of the Northeast Monsoon
Tropical	Continental	NT_I, NT_T	North India, Tibet	Reaches Malaya as upper westerlies during the Northeast Monsoon
		NT_A	Australia	Reaches Malaya as southerlies, during the Southwest Monsoon and may be confused with subsided T_{SI}
Tropical	Maritime	T_{NP}	North Pacific	Reaches Malaya as an extension of the Northeast Trades during the Northeast Monsoon
		T_{SI}	South Indian Ocean	Reaches Malaya as an extension of the Southeast Trades during the Southwest Monsoon
		$NT_{SP\ \{equator\}}$	South Pacific	Reaches Malaya as easterlies modified along the equator E_M

Horizontal Air-stream Analysis

The methods of air-mass analysis used in temperate latitudes are not at present applicable to equatorial regions, where another technique has been developed. The source regions of the air masses are fairly well-known and, ignoring fine distinctions which may represent only subdivisions of the main air streams, generally not more than three streams are present in the region at the same time. Thus, given a good network of upper-wind reports, the flow in each stream may be traced from the source to the Equator and even into the other hemisphere. (Examples applying this technique are in Chapter XII.) Recognition and tracing of the streams are facilitated by the few streams involved in equatorial latitudes.

Tracing the streams from source is implicit in temperate latitude air-mass analysis, but the equatorial analyst discards all the other methods of temperate air-mass analysis so that his is aptly described as 'Air-stream Analysis'. Upper-wind observations from a number of stations and from aircraft are plotted on maps, each giving conditions at one selected level for one reporting hour. Stream-lines are sketched in the direction of flow and spaced inversely propor-

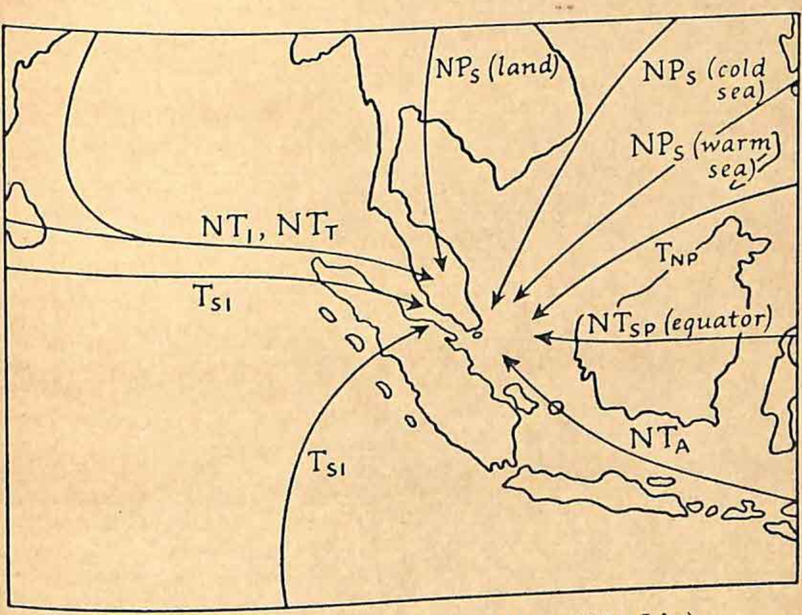


FIG. 49 Principal Air Streams Affecting Malaya (After John)

tional to wind speed on a scale suiting the map;* e.g. for strong winds the stream-lines are close together. Considerable smoothing is necessary for the finished stream-lines, and an example of an analysed stream-line chart is shown in Fig. 50.

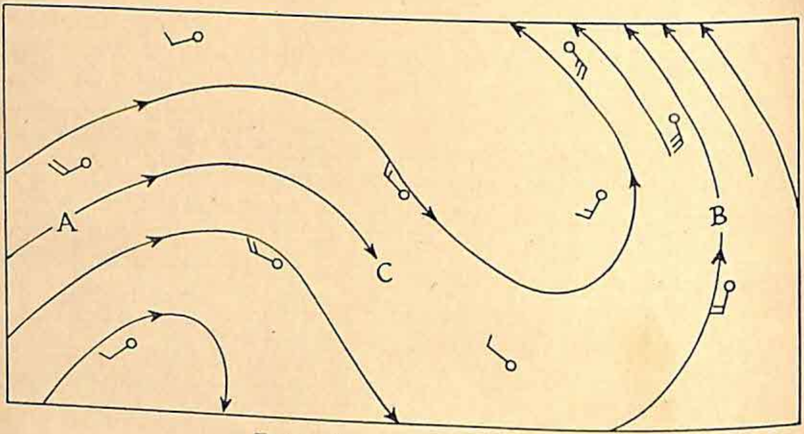


FIG. 50 Stream-line Drawing

The correct drawing of stream-lines requires experience and previous knowledge of the types of streams prevailing. Interpreting them follows these rules:

(1) Parallel stream-lines means there is neither convergence nor divergence in the current (A of Fig. 50).

(2) Where a stream-line begins in an area of parallel flow there is divergence, because increasing velocities denote that more air is leaving the region than is entering it. Compensation for this outflow must come by air subsiding from higher levels (B of Fig. 50).

(3) Where a stream-line vanishes in an area of parallel flow (C) there is convergence, because decreasing velocities denote that more air is entering the region than is leaving it horizontally. The excess air must be rising, and favours the formation of clouds and precipitation.

(4) Where the spacing increases, the directions indicate divergence, though the decreasing speeds indicate a compensating convergence, and the net effect is likely to be small.

* C. E. Palmer (in *Quarterly Journal of Royal Meteorological Society*, Vol. 78, No. 336) describes a different method of stream-line drawing which might have some advantages over that described herein if the density of the observational network were greatly increased. In Palmer's method (adapted from Bjerknes) stream-lines are continuous and spacing is independent of wind strength, the latter being described by separate lines of equal wind strength, known as 'isovels' or 'isotachs'.

(5) Where the spacing between the stream-lines narrows, wind directions indicate convergence, though the increasing speeds indicate divergence with small net effects.

Where there is a good network of upper-wind observations, these rules are sound in application, making it possible to forecast cloud development following the recognition of areas of convergence and divergence within a stream. But where the network of upper-wind observations is sparse, the meteorologist may be compelled to draw smoothed stream-lines regardless of spacing. Estimates of convergence and divergence within a stream may not then be reliable, yet the streams are sufficiently indicated so that, where two meet, identifying them is possible from their previous history.

3. Air-stream Boundaries

The line separating two streams of similar properties in equatorial regions might well be known as an 'Air-stream Boundary'. It is analogous to the 'Air-mass Boundary' of temperate latitudes, but the latter implies, from theoretical background and from usage, the existence of identifiably different properties in the two streams. Such differences do not exist near the Equator other than in the directions, so that the boundary is evident on the upper-air charts solely from the meeting of the air streams.

As the geostrophic equation does not apply close to the Equator, two streams do not necessarily shear in a cyclonic sense, and the streams often turn to become asymptotic to their boundary. The practice in drawing stream-lines is, in the absence of contrary observational evidence, to bend the stream-line to approach the boundary asymptotically in the same sense as the strongest nearby wind observation.

The Surface Air-stream Boundary

Surface air-stream boundaries cannot be demarcated as accurately as those of the upper layers. The analyst attempts to find them from the surface winds, but at low latitudes these are greatly influenced by local effects. Winds at all levels are frequently light, and land breezes and sea breezes may mask the main flow at lower levels. This is illustrated in Fig. 51 by hourly surface-wind observations taken at Penang on 6th January 1951, at a period when the North-east Monsoon could be traced aloft across Malaya and the Straits of Malacca.

Over the oceans, surface-wind observations from ships and small islands may be considered sufficiently representative of the main

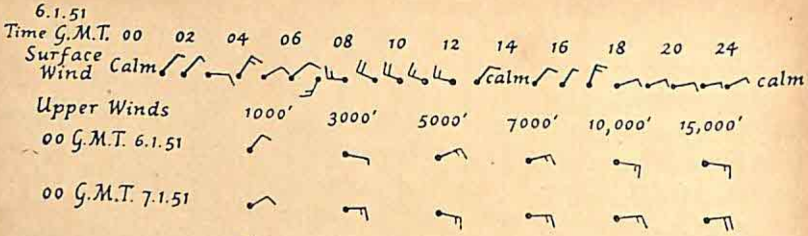


FIG. 51 Surface Winds at Penang

air stream to mark the position of the surface boundary. A coastal wind may be accepted as similarly representative if its direction is contrary to the land or sea breeze for that time of day and providing there is no evidence of a local storm. Thus, an easterly to north-easterly surface wind at Malacca in the late morning during January would be evidence, in the absence of a thunderstorm or cumulonimbus, that the surface boundary of the Northeast Monsoon lies to the west.

Because it is rare that much reliability may be attached to evidence of surface winds over the major land masses, three auxiliary methods are used for obtaining a second approximation for the position of the surface air-stream boundary. These methods are:

(1) Drawing the surface boundary through places where there is excessive precipitation or cumuliform development, particularly when these occur at times of day when conditions are unfavourable for convection.

(2) Taking the boundary at some low upper-level (1000 to 3000 ft., for instance) and assuming that the surface boundary is vertically below this.

(3) Placing the boundary in a pressure trough.

To assume that excessive cloud development or precipitation marks the surface air-stream boundary is justifiable on the grounds that excessive cloud development must occur where convergence is greatest, and greatest convergence is to be expected at the junction of two streams. If the rising air within the clouds has originated at an upper level, it might be assumed that surface air was later drawn into the cloud system, because reports of cumulonimbus with base above 3000 ft. are rare. In turn, if the slope of the surface through the air-stream boundaries at successive levels were not great, the drawing of surface air into the base of the cloud system would constitute a movement of part of the surface boundary to a position below the cloud base.

Whether or not the cloud development (and thus the surface air-stream boundary) may be represented by a line on the chart is disputed, and some writers prefer to treat development as scattered within a zone.[59] It is practical to consider the line as the nearest approximation to the surface air-stream boundary, and as denoting a mean position to which the boundary tends to return after any distortion. It represents the mean convergence position of the two streams, and consequently of maximum cloud development.

Considering the surface boundary to be in a vertical plane with the boundary at a slightly higher level introduces an error, because the surface through the boundaries at various levels frequently slopes.[57] If it is correct that major cloud development is usually above the surface boundary, the method gives a reasonable approximation, because Mather[60] has shown that the location of maximum precipitation correlates with the position of the air-stream boundary at the 3000-ft. level.

Opinions differ about placing the surface boundary in the isobaric trough, and there are good arguments for and against the method. Of about seven hundred charts examined during 1948 and 1949, none showed an air-stream boundary coinciding with a region of high pressure, and in each case a trough could conveniently be drawn without undue distortion of the 1-millibar isobars, along the line of the surface air-stream boundary as located by air-stream analysis.

The air-stream boundary of the surface synoptic chart then, though fundamentally for demarcating the junction of the surface streams, is difficult to determine because some sections of it are located by cloud and precipitation, others by winds above the ground level, some by the pressure trough and many by a combination of these signs.

Moving Air-stream Boundaries

The boundaries at the various levels are determined in the first place by discontinuities in the stream-lines at a particular time (the instantaneous stream-lines). This method alone is not always reliable and, to achieve continuity from one day's weather chart to the next, the trajectories in each stream over a period must be considered. Consider first the stationary boundary in Fig. 52*a*. Suppose it undergoes a horizontal displacement in a direction transverse to its length. From the time of the displacement the air ahead of the displacement must retreat, and typical trajectories will be as in Fig. 52*b*.

If an analysis of the instantaneous stream-lines had been made

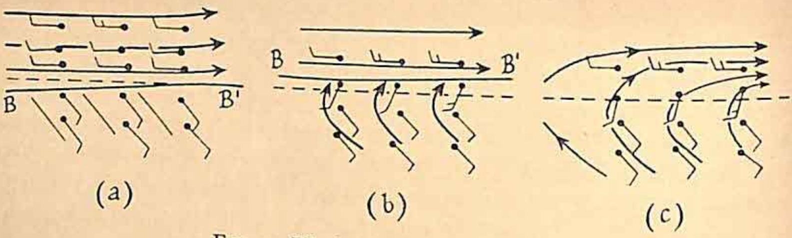


FIG. 52 Displacement of a Boundary

for the actual observations spaced as in Fig. 52*b*, the pattern would (Fig. 52*c*) either lose trace of the boundary or show a displacement of it unduly great in relation to the actual winds. This difficulty is avoided if, when a boundary is undergoing horizontal displacement and its changing location is not shown by the instantaneous stream-lines, it be drawn in the position where it might be expected to have travelled under the influence of the observed winds. To this extent the air-stream analysis is not solely an analysis of the instantaneous stream-lines.

4. Intertropical Front

Air-mass boundaries of middle latitudes are sometimes termed 'fronts'. Many writers use 'Intertropical Front' for the boundary between streams from different hemispheres—to which there are objections. There is never a true air-mass discontinuity or 'front' at low latitudes because the vertical structure of the streams is similar, and, far from being a simple 'front', the meeting-place consists of a system of one or two equatorial boundaries. Three streams are frequently involved—exemplified in a typical weather chart for October 1947 (Fig. 77). On this chart are two boundaries with Equatorial Westerlies streaming between them. To the north of the Northern Boundary is the Northeast Monsoon, and to the south of the other are the Southeast Trades. The orientation of the Equatorial Westerlies varies frequently between west-southwest and west-northwest, increasing the convergence on one boundary and decreasing it on the other. Because the Westerlies are roughly along the Equator, each boundary in turn is the 'boundary between streams from different hemispheres' (or Intertropical Front), and at the same time an active convergence line. The employment in equatorial latitudes of the term 'Intertropical Front' leads to the impossible position of having a 'front' which moves hundreds of miles from day to day with periods, if the westerlies are oriented from the west, when neither boundary fulfils the definition of an

'Intertropical Front'. Crossley[47] avoids the term and writes of the 'Intertropical Convergence Zone'.

It is best to abandon the use of the term 'Intertropical Front' between latitudes 10° N. and 10° S. and to replace it by the expression 'Equatorial Air-stream Boundaries'. Outside these latitudes, however, the term is permissible because contrasts may occur in the vertical structures and rarely more than two streams are involved.

CHAPTER IX

Slope of Air-stream Boundaries

In this chapter the mathematical aspect of air-stream boundaries will be examined.

1. The Equation of Slope

The angle of inclination (Fig. 53) of a surface of discontinuity between two air currents of different temperatures and velocities may be expressed[61] as:

$$\tan \alpha = -\frac{2\omega \sin \phi(u_1 T_2 - u_2 T_1)}{g(T_2 - T_1) - 2\omega \cos \phi \cos \beta(u_1 T_2 - u_2 T_1)} \quad \ldots \quad (1)$$

where α is the angle of slope of the surface of discontinuity;

- ω the angular velocity of the earth;
- ϕ the latitude;
- g the gravitational constant;
- β the angle between the intersection of the surface with the ground, and the east–west line;
- T_1 and T_2 are the temperatures on each side of the intersection;
- u_1 and u_2 are the velocities in each current parallel to the intersection.

The assumptions in this formula are (i) that differences in density are solely due to differences in temperature, and (ii) that there is steady horizontal motion. The former is understandable because the effect of temperature should outweigh other factors affecting density. The latter narrows the application of the equation because the size of accelerations at low latitudes has not yet been satisfactorily investigated.

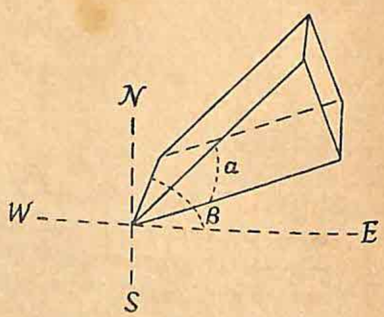


FIG. 53 Slope of Surface of Discontinuity

When there is No Temperature Difference

The temperature difference between two equatorial streams is very small,[8] and if it is treated as nil, equation (1) may be simplified to show the slope for a surface of discontinuity in velocity alone (i.e. when $T_1 = T_2$):

$$\tan \alpha = \frac{\tan \phi}{\cos \beta} \quad . \quad . \quad . \quad . \quad (2)$$

Thus for surfaces of discontinuity running in an east–west direction, when

$$\beta = 0,$$
$$\alpha = \phi \, . \quad . \quad . \quad . \quad . \quad (3)$$

Thus when a surface of discontinuity is oriented east–west and there is no difference of temperature, the angle of slope of the surface of discontinuity equals the latitude. For example, at the Equator, slopes will be horizontal: at 5° N. and 5° S. they will be 1 in 9, and at 10° N. and 10° S. 1 in 6.

When the surface of discontinuity is oriented north–south, $\beta = 90°$, and $\cos \beta = 0$, so that for finite values of ϕ slope is vertical. At the Equator the value of α becomes indeterminate, both $\tan \phi$ and $\cos \beta$ being zero.

Thus, if the discontinuity is one of velocity alone, slopes may vary from vertical to horizontal, and small variations in orientation or in latitude produce large variations in slope. John[8] has discovered no appreciable differences of temperature between currents at low latitudes, yet probably small real differences do exist since the air-craft reports used by John are rarely accurate to within 1° F., and larger errors must occur. For that reason it is necessary to examine the formula for slope when small temperature differences exist each side of the discontinuity.

When there is a Temperature Difference of 1° F. or Less

Equation (1) is frequently stated as Margules' Equation, i.e.

$$\tan \alpha = \frac{2\omega \sin \phi}{g} \cdot \frac{T(u_1 - u_2)}{T_1 - T_2} \quad . \quad . \quad . \quad (4)$$

This is obtained:

(*a*) By assuming that the temperatures are similar and that no great error is introduced by substituting $T(u_1 - u_2)$ for $(u_1T_2 - u_2T_1)$, where T is the mean of the temperatures: the assumption is reasonable because at low latitudes, temperatures are of the same order.[8]

(*b*) By discarding the second term in the denominator of Equation (1) as negligible relative to the first. This is justifiable because if the discontinuity lies in a north–south direction, then $\beta = 90°$ and $\cos \beta = 0$, so that $2\omega \cos \phi \cos \beta (u_1 T_2 - u_2 T_1)$ is zero for all latitudes.

On the other hand, if the slope is oriented east–west, $\beta = 0$ and $\cos \beta = 1$. Then, assuming a mean temperature of 75° F. (296·9° A), the second term $2\omega \cos \phi \cos \beta . T(u_1 - u_2)$ for various values of latitude ($\phi = 1$ to 10) and various values of wind shear ($(u_1 - u_2) =$ 1 to 15 m.p.h.) becomes as in this table:

$u_1 - u_2$	$\phi = 1$	$\phi = 5$	$\phi = 10$
1 m.p.h.	1·9	1·9	1·9
5 ,,	9·6	9·6	9·5
10 ,,	19·3	19·2	19·0
15 ,,	28·9	28·8	28·5

The value of the first term in the denominator (i.e. $g(T_2 - T_1)$) is 545 if we assume a temperature difference of 1° F. For a $(u_1 - u_2)$ of 1 m.p.h., the second term is 0·3 per cent. of the first.

If $(u_1 - u_2)$ is 5 m.p.h., the second term is 1·8% of the first.
,, ,, ,, 10 ,, ,, ,, ,, ,, 3·5% ,, ,, ,,
,, ,, ,, 15 ,, ,, ,, ,, ,, 5·3% ,, ,, ,,

As winds in any equatorial current are most frequently about 10 m.p.h., Margules' Equation appears sufficiently accurate for the types of slopes near the Equator. The error decreases at latitudes away from the Equator.

Let us now estimate the slope in various equatorial latitudes for varying values of $(u_1 - u_2)$, assuming a mean temperature of 75° F. and a difference of 1° F. between the temperatures of the currents. The slopes plotted in Fig. 54 were computed by Margules' Equation and thus apply exactly for a north–south orientation. The main points brought out by comparing these curves are that: (1) slopes are gentler near the Equator for the same values of $(u_1 - u_2)$; (2) at latitudes 5° to 10°, small variations of $(u_1 - u_2)$ are associated with large variations of slope when $(u_1 - u_2)$ is less than about 5 m.p.h.; (3) when $(u_1 - u_2)$ is large between latitudes 5° and 10° slopes are steep and do not vary greatly.

Thus we can get a theoretical picture of the slope of a surface of

discontinuity at low latitudes. Lacking accurate data on the dimensions involved, we have first taken the case where both currents are of the same temperature. Then slopes vary from vertical (when the discontinuity is oriented north–south) to a gentler slope for an east–west orientation. The gentler slopes range from horizontal at the Equator to 1 in 6 at latitudes 10° N. and S.

If the temperature difference is real and not greatly exceeding 1° F., slopes vary greatly but orientation is not important. Within the likely range of wind shear at the discontinuity, slopes are all gentle and diminish as the shear diminishes.

The slope of the Northern and Southern Equatorial Air-stream Boundaries* (two known surfaces of discontinuity in the Malayan–East Indies area) were measured[57] each day between 21st July 1947 and 20th July 1948. The observations of slope on both the surfaces of discontinuity may be tabulated in order of frequency as follows:

Slope from Ground to 10,000 ft.

Slope	Vert. to 1 in 49	1 in 50 to 1 in 99	1 in 100 to 1 in 149	1 in 150 to 1 in 199	1 in 200 to 1 in 249	1 in 250 to 1 in 299	1 in 300 to 1 in 349
No. of Occasions	74	54	27	23	18	12	14

Slope	1 in 350 to 1 in 399	1 in 400 to 1 in 449	1 in 450 to 1 in 499	1 in 500 to 1 in 549	1 in 550 to 1 in 599	1 in 600 and Greater	Total No. of Obs.
No. of Occasions	17	5	4	9	1	7	265

These slopes, decided on the wind-field, were measured at right angles to the ground-level discontinuity. Orientations were disregarded, but were mostly northeast to southwest. Because the observations were few, it was not possible to determine whether there was steady horizontal motion, but if Margules' Equation is applicable, several interesting comparisons arise. In the first place, 28 per cent. of the observations fall within the vertical to 1/49 range,

* A description will be found in Chapter XI. The Northern and Southern Air-stream Boundaries are sometimes known as the Northern and Southern Convergence Lines.

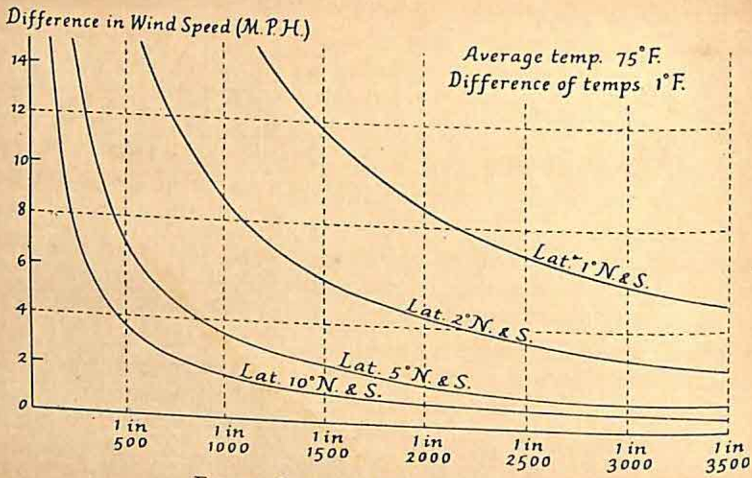


FIG. 54 Computed Slope (Margules)

possibly accounted for by discontinuity of wind without discontinuity of temperature; 10 per cent. of the observations have slope of 1/400 or less, possibly associated with temperature differences of about 1° F. (Fig. 54).

Of the observations, 62 per cent. were slopes from 1/50 to 1/399 and most probably associated with temperature differences less than 1° F. To verify this let us consider the slopes of different discontinuities at lat. 5° when $(u_1 - u_2) = 5$ and $T = 296{\cdot}9°$ A computing from Equation (1).

Type of Discontinuity	*Orientation of Discontinuity*	
	East–West	*North–South*
Wind alone . .	1/9	Vertical
½° F. Temp. Difference .	1/340	1/350
1° F. Temp. Difference .	1/690	1/700

From this calculated table, most of the slopes observed could have been associated with small differences of temperature. The magnitude of the differences would be less than 1° F. and frequently less than ½° F. As the present state of meteorological instrumental accuracy does not permit recognition of temperature differences of this order, the slope formula is impractical at present.

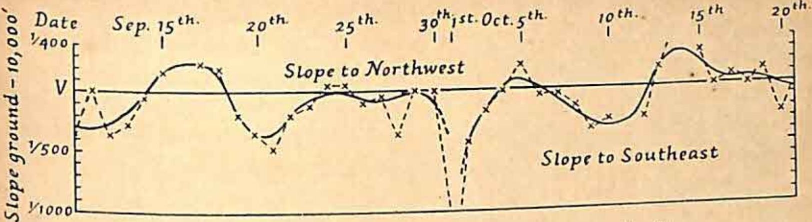


FIG. 55 Changes of Slope of a Surface of Discontinuity

Improved accuracy in measurement may some day disclose such small temperature differences, but it will be difficult in any case to decide whether such differences are of the streams or due to local convection. Discontinuities ('air-stream boundaries') are commonly associated with convection and, because the mean diurnal surface temperature range at equatorial coastal stations is between 10° and 15° F., convection must bring about appreciable local changes in temperatures at upper levels.

If convectional effects can locally influence the temperature distribution in the upper air, the slope of a surface of discontinuity might be expected to fluctuate haphazardly in size and sense, but observations[57] do not support the idea. Fig. 55 shows the slopes of one discontinuity in the Malayan–East Indies area, measured from the wind-field at 0000 hours G.M.T. (0730 Malayan time) each day from the 10th September to 21st October 1947. At times slope fluctuated within a short period, either through convection or through errors of measurement. On the other hand, there was some regularity in the way the slope changed slowly from one sense to another over longer periods. The differences of temperature (and therefore of density) were apparently real and possibly measurable.

2. Slopes in Equatorial Southeast Asia

From 46 observations during 1947 and 1948,[57] the mean slope of the Northern Equatorial Air-stream Boundary between ground and 5000 ft. was about 1 in 100; over the same period, 44 observations showed the mean slope between 5000 and 10,000 ft. to be 1 in 75. The mean of 74 observations between ground and 10,000 ft. gave a slope of 1 in 70. Slopes to altitudes greater than 10,000 ft. were not investigated, but from limited data available at the time, higher-level slopes appeared to be nearly horizontal.

The mean slope from the ground to 10,000 ft. was nearer vertical than in either of the two component layers, though the difference lay within the range of observational error. It was partly occasioned by the occurrence of slope reversal with altitude; e.g. on 27th

October 1947; slope from ground to 5000 ft. was 1 in 220; slope from 5000 ft. to 10,000 ft. was 1 in 120 in the opposite direction and the effective slope from ground to 10,000 ft. was 1 in 50. Four cases of slope reversal were included in the 74 observations.

In few cases were slopes nearer the horizontal than 1 in 200; their mean values lay midway between those accepted for cold and warm fronts in temperate latitudes—1 in 50 and 1 in 150 respectively.[2] Tests showed that a seasonal variation was improbable.

Extending this analysis to the Southern Equatorial Air-stream Boundary showed[57] that slopes there were less than on the Northern Boundary. The mean slope of the layer from ground-level to 5000 ft. was 1 in 220, and for the layer from 5000 to 10,000 ft. the slope was only 1 in 240. Slope from ground to 10,000 ft. (from 191 observations) was 1 in 210, and there were eighteen cases of slope reversal with height.

The slope of the Southern Boundary was slight and frequently less than most temperate-zone warm fronts. Departures from the mean were greater than those of the Northern Boundary. The slope of the Southern Boundary appeared always in a state of change whereby the sense of the slope was periodically reversed, so that no mean could describe it adequately. No relation could be found between the degree of slope and the orientation of the boundary, but some slight seasonal character was discernible in June, July and August, when mean slopes were slight.

Slope of Moving Air-stream Boundaries

Air-stream boundaries undergoing horizontal displacement have also been observed. Because observations were few, both Northern and Southern Boundaries were treated together. They were arranged in two classes—'Overriding' and 'Undercutting and Vertical'—in order to clarify changes of slope dependent on the direction of movement. The division was artificial, because boundary surfaces nearly vertical in the layer up to 10,000 ft. were included as cases of undercutting (i.e. cases where the slope was back from the direction of travel as in a temperate-zone cold front). Mean slopes were:

Layer	0–5000 ft.	5000–10,000 ft.	0–10,000 ft.
Undercutting and Vertical .	1 in 110	1 in 140	1 in 130
Overriding . . .	1 in 240	1 in 270	1 in 270

There was only slight difference of slope between layers when movement entailed undercutting, and it appears that surface retardation did not appreciably steepen slopes in the lowest layer, as occurs in a temperate cold front. Similarly, variations of slope within one layer in cases of overriding were too small to permit assuming that lower slopes were flattened. Thus it may be assumed that movement does not affect slope in the lowest layer.

A marked difference existed, throughout the three thickness ranges, between slopes of undercutting and those of overriding. Excluding cases of near vertical slope, the mean slope of undercutting between ground and 10,000 ft. was 1 in 210—appreciably greater than the 1 in 270 of overriding. It appears, then, that up to 10,000 ft. the slope of the Equatorial Air-stream Boundaries is steepened by undercutting and reduced by overriding.

While little information is available concerning the slope of air-stream boundaries in other equatorial regions, the following implications may be drawn from the observations in the Malayan–East Indies area:

(1) Slope of stationary Equatorial Air-stream Boundaries below 10,000 ft. may lie either side of the vertical, and range from 1 in 70 to 1 in 240 (about 1° or less to the horizontal).

(2) Slope above 10,000 ft. is generally nearer the horizontal.

(3) Slope of a moving boundary is slightly steepened up to 10,000 ft., when the movement involves undercutting, and slightly flattened, when one stream overrides another.

CHAPTER X

Cloud Structure on Air-stream Boundaries and Convergence Lines

Equatorial Air-stream Boundaries, commonly located in areas of convergence, are associated with vertical air motion which favours cloud formation. Aircraft encounter long lines of cumulonimbus or banks of altostratus near the boundaries, a condition resembling that on fronts of temperate latitudes. The cloud-lines are not always uniformly spread along the surface air-stream boundary, but a concentration of cloud is generally found near it, which accounts for the tendency to consider cloud structures at equatorial boundaries as similar to those of higher latitudes. The falsity of this view can be illustrated by comparing the conditions involved.

1. Cloud Structure on Fronts of Temperate Zone

The term 'front' describes the boundary between warm and cold air. A technique of frontal analysis is general throughout the temperate regions and polar latitudes.[62] How far it is applicable to equatorial or tropical regions is the subject of conjecture.

The frontal theory is discussed in standard meteorological texts so that only an outline of it is needed before discussing equatorial conditions.

A 'front' is a line where the bounding surface dividing two air masses meets the ground. It usually lies in a trough of low pressure with isobars turning sharply in a cyclonic sense across it; if convergence is strong, cloud and precipitation line the front. Sometimes a front may lie parallel to the isobars when, if convergence is weak, its edges will be diffuse.

When the front advances so that the cold mass is replacing the warmer, the air-mass boundary at the ground is termed a 'Cold Front'. The colder air undercuts the warmer, and the front is marked by cumuliform cloud and by rain in a zone which may be about 50 miles broad. The cloud form varies with the degree of instability and of convergence, though cumulonimbus is common. When the cold front passes over an observing station, the wind veers

(in the Northern Hemisphere), temperatures fall, pressure begins to rise and a line squall may occur.

When the warm air mass is replacing the colder, the boundary is known as a 'Warm Front'. The warmer air rides over the colder ahead of the air-mass boundary at the ground, and a shelving layer of stratiform cloud develops on the sloping surface. At high levels, as far as 600 miles horizontally ahead of the boundary on the ground, cloud is cirrostratus, gradually lowering to altostratus and nimbostratus as the front advances. Rain begins as much as 200 miles ahead of the front, and there is a rise of temperature when it passes over a station. Winds frequently veer a little (in the Northern Hemisphere) and the barometer occasionally rises or steadies. Cross-sections of fronts are shown in Fig. 56.

There are resemblances between the cloud systems of fronts and of Equatorial Air-stream Boundaries; in each, cloud forms on the boundary between air masses or streams of different origin. On the other hand, fundamental differences exist. The streams do not necessarily shear cyclonically either side of an equatorial boundary (as they do about a temperate front) owing to the negligible geostrophic force near the Equator. Furthermore, the theory of temperate fronts is based on physical differences (mainly of temperature and density) in the opposing masses, whereas such differences in equatorial air masses are small.

A refinement in the theory of fronts is the 'Occlusion'. Fig. 57*c* shows the temperate-latitude chart to which the cross-sections of Fig. 56 relate. AE is a cold front, and AFB a warm front in Fig. 57*c*. The cross-sections of Fig. 56 are on the vertical plane through CXD of Fig. 57*c*. On the frontal system of Fig. 57*c*, a wave depression is forming, and winds are blowing clockwise around it (in the Southern Hemisphere). Near its centre, the easterly movement of the warm front is less than that of the cold front. Eventually the section AE of the cold front and AF of the warm front coincide and later AE passes AF to form an 'Occlusion', which means that the portion EAF of the 'warm sector' no longer exists at ground-level, but persists aloft, bounded by the sloping surfaces of the two fronts.

Occlusion may take place in one of the two ways illustrated in Fig. 58*b* and *c*. If the air following the cold front is colder and denser than that preceding the warm front, it undercuts both warm front and warm sector (Fig. 58*b*). If the air following the cold front is less cold than that preceding the warm front, both warm sector and cold front override the preceding air by moving up the slope of the warm front (Fig. 58*c*). In the intermediate case,

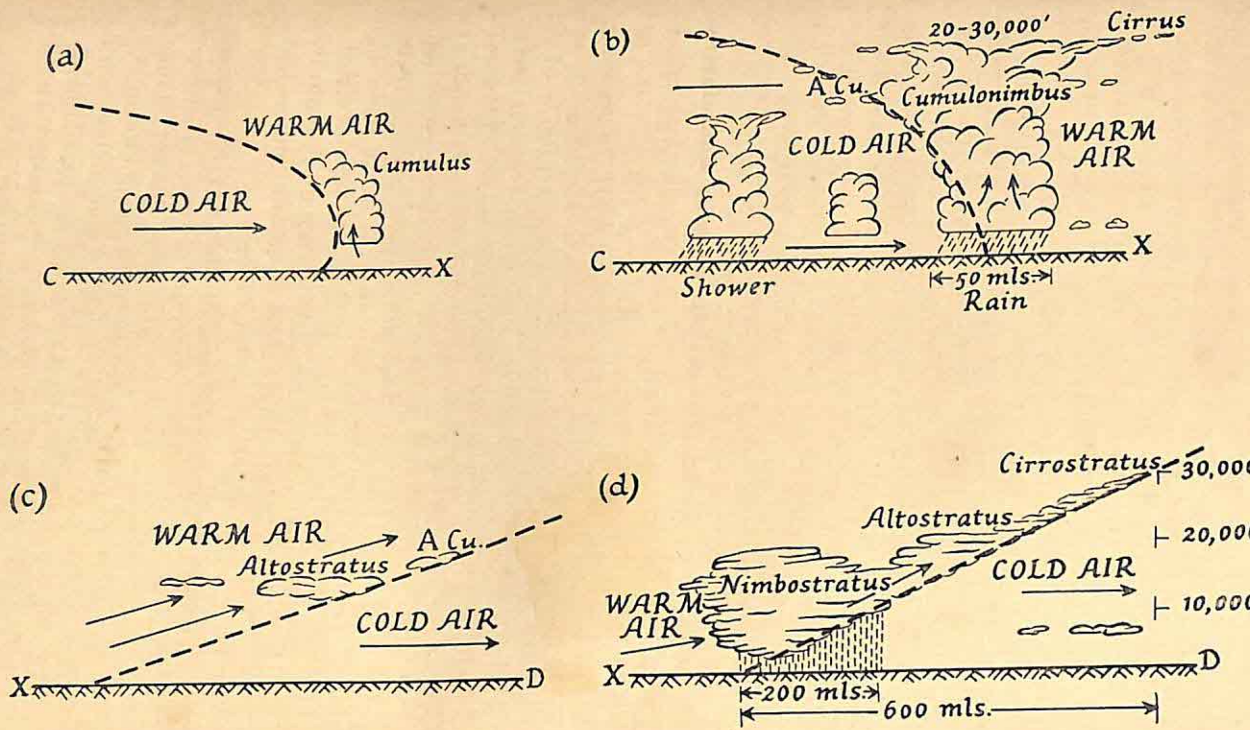


FIG. 56 Fronts of Temperate Latitudes

(*a*) Weak cold front (*b*) Vigorous cold front (*c*) Weak warm front (*d*) Intense warm front

when the temperature in the air before the warm front is nearly the same as that behind the cold one, the front at the ground is non-existent and cloud is restricted to altostratus and cirrostratus.

The cloud resulting from an occlusion is usually a combination of cumuliform and stratiform, therein resembling the lines of cloud common on the equatorial air-stream boundaries, though their geneses are different. The formation in an occlusion depends on the presence of two streams of contrasting temperatures which are not present in equatorial latitudes. Furthermore, the occluding process is associated with a cyclonic circulation—which is rare near the Equator, where the geostrophic force is small and winds cross the isobars at a large angle. The occlusion theory thus cannot be used to explain the cloud-lines at equatorial boundaries. The points of resemblance between equatorial boundaries and temperate fronts need further investigation, because the former have slope despite the lack of evidence about temperature contrasts, and because cloud on the equatorial boundaries may be cumuliform or stratiform.

2. Cloud Formation on Equatorial Air-stream Boundaries

An analysis[57] was made from July 1947 to July 1948 of the cloud structure of the Northern and Southern Air-stream Boundaries of Southeast Asia located by the methods of Chapter VIII. In deciding the representative cloud types each day, peculiarly local cases were carefully excluded. The types reported were representative of a considerable section of the boundary, and diurnal effects were excluded after comparing clouds over both land and sea. This was necessary because the diurnal variation of cloud over equatorial seas is greater than in temperate latitudes. During the investigation many cases occurred in which, although in the early morning there was a bank of towering cumulus at sea, clear skies lay over the landward portion of the same boundary. It was usual for this cloud-line over the sea to be obliterated or reduced to cumulus humilis by the afternoon, while a well-developed bank of cumulonimbus had developed on the landward extension of the boundary.

Sheets of altostratus were not affected in that way by diurnal changes. Extensive sheets were often associated with an air-stream boundary, normally persisting for days with little diurnal change other than thinning or temporarily dividing to altocumulus in the evening. Smaller sheets of altostratus found in conjunction with isolated cumulonimbus clouds over land generally dissipated during the evening. No record was made of altostratus associated with a boundary unless the sheet was extensive and persistent.

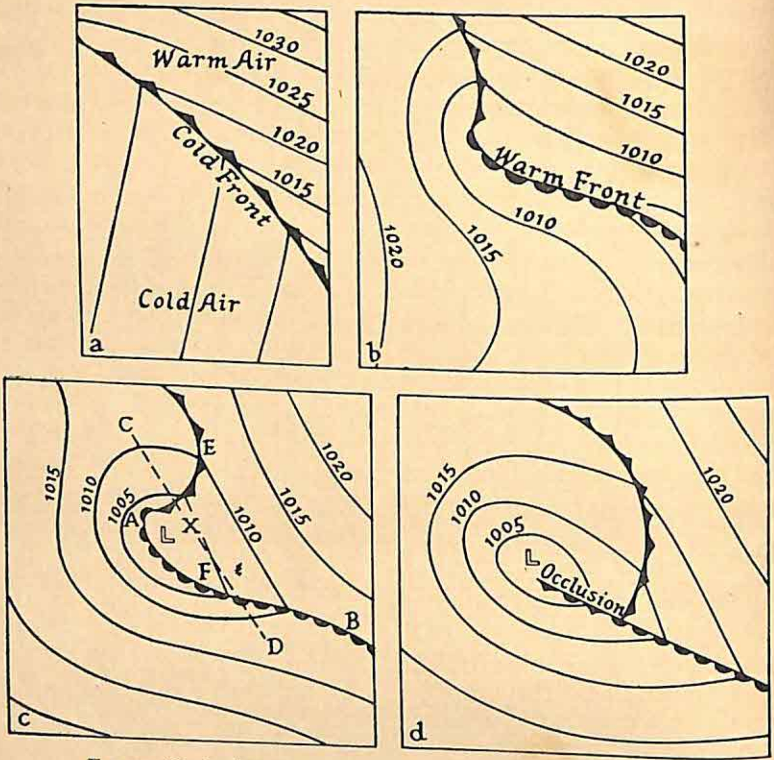


FIG. 57 Occlusion on Synoptic Chart—Southern Hemisphere

(*a*) Cold front
(*b*) Wave forming
(*c*) Open wave
(*d*) Partly occluded wave

In tabulating the results, the frequency of distribution of cumuliform and stratiform clouds was found to be nearly equal on each of the boundaries:

Cloud Type	*Percentage of Observations*		
	Northern Boundary	*Southern Boundary*	*Combined Boundaries*
Cumuliform solely	33	35	34
Stratiform solely	21	20	21
Cumuliform with Stratiform . . .	33	21	25
No excessive cloud	13	24	20

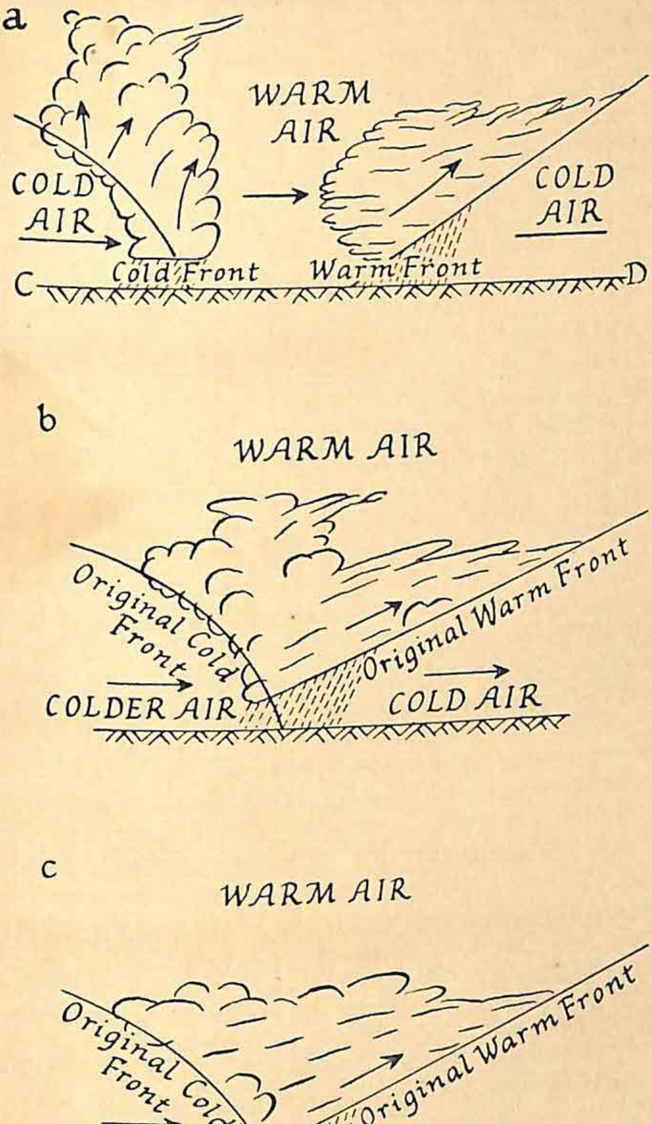


FIG. 58 Occlusion Cross-sections

(*a*) Cross section along CD of diagram 57*c*
(*b*) 'Cold' type occlusion
(*c*) 'Warm' type occlusion

Cumuliform was present in 59 per cent. of all observations and stratiform on 46 per cent. Thus cumulus and stratus formations, singly or combined, occurred with comparable frequencies on the boundaries. The long-period approach concealed seasonal differences. Occasions when the ratio of stratiform to cumuliform clouds was rising coincided with decreased convergence on the boundary, suggesting that the altostratus may have been formed by shear as well as by spreading cumulus tops. These facts do not exclude the possibility that altostratus may be formed by upslide on the air-stream boundary as in a temperate warm front, for which reason the relation of the altostratus sheet to the slope of the air-stream boundary was investigated.

The average height of the altostratus in the region was taken to be about 12,000 ft., but slope at that level was not measured because upper-wind observations were not plentiful. No cases of slope reversal were observed from 10,000 to 15,000 ft., and thus using the ground to 10,000-ft. slope was considered justified.

Neglecting cases of vertical slope, occasions when altostratus formed on only one side of the air-stream boundary were more frequent than when on both in the ratio 82 : 28 or about three times as often. This might indicate that altostratus formed on the slope of the boundary, except that in a separate investigation of cases of vertical slope the altostratus occurred on only one side or on both with equal frequency.

On occasions when the cloud sheet was on only one side of the ground boundary, 39 per cent. had cloud on the side of the boundary opposite to the slope (i.e. wholly within one of the streams). When the cloud sheet occurred on the sloping surface, 55 per cent. occasions showed altostratus associated with towering cumulus or cumulonimbus, pointing to a genesis in spreading cumulus tops rather than in upslide.

Examining the cloud types in relation to the slope of moving boundaries, it was found that in cases of overriding the frequency of altostratus equalled that of cumuliform cloud, while in cases of undercutting the ratio of stratiform to cumuliform was 5 : 9. The frequency of altostratus on an overriding slope nearly equalled the frequency of altostratus on the side of the ground boundary opposite the overriding slope. Altostratus sheets on the slope of an undercutting air-stream boundary were four times more frequent than opposite the slope.

This gave no evidence that upslide altostratus forms on an air-stream boundary (as in a temperate-latitude warm front), while indicating that most altostratus is formed in the shear associated

with spreading tops of cumulus. A 'warm front' theory was not positively excluded, but if warm-front conditions occur at all near the Equator, they are rare. It is necessary, then, to consider the equatorial boundaries from a new aspect.

3. Structure of Vertical Air-stream Boundaries

If we disregard the effects of the diurnal temperature variation on cloud growth, the amount of cumulus and cumulonimbus on the boundary is related in some way to the convergence of the two air streams. Because winds in equatorial latitudes are rarely strong, the convergence is slight. Concentrations of up to 4/8 towering cumulus and cumulonimbus along or near the boundary occur without abnormal convergence, and the rising currents in the clouds are sufficient to discharge upwards such air as is converging on the boundary, so that concentrations of cloud farther from it do not develop.

Consider a zone along a vertical air-stream boundary of length x miles and width y miles, and suppose that, from ground to altitude h ft., the opposing moist streams are of average horizontal velocities u_1 and u_2 m.p.h. at angles θ_1 and θ_2 respectively to the boundary (Fig. 59).

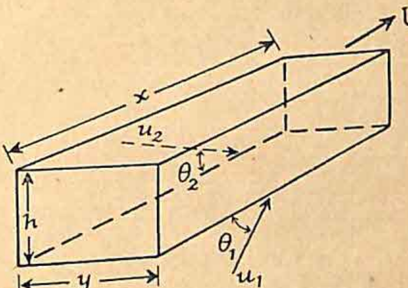


FIG. 59 Convergence at an Air-stream Boundary

Let U be the mean horizontal outflow from the downstream end of the zone.

Taking sample values which represent common conditions in equatorial regions, let $u_1 = u_2 = U = 10$ m.p.h., and let $\theta_1 = \theta_2 = 45°$. Consider the condition when $h = 10{,}000$ ft. Assuming that pressure and temperature are unchanging, the amount of air rising above 10,000 ft. is for various values of x and y:

$x =$	100	500	1000 mls.
$y = 10$	2640	13,440	27,940 cu. m.p.h.
50	2400	13,350	26,700 ,, ,,
100	2100	12,900	26,400 ,, ,,
300	900	11,700	25,200 ,, ,,

If it be assumed that air ascending in the zone is confined to

cumulonimbus development, by taking an average up-current within the cloud-columns an estimate may be made regarding the amount of cumulonimbus in the zone. There have been many different appraisals of the up-currents in cumulonimbus clouds; Piercey[63] writes that upward gusts of 25 ft. per sec. are not unusual, but it is probable that the average up-current over a great depth of atmosphere is less. Measurements in cumulonimbus were made[17] in flights over Florida and Ohio in 1946 and 1947, where mean up-draughts at various levels were found to be:

Ohio . .	At 5000 ft.	Up-draught	14 ft./sec.
	At 10,000 ,,	,,	21 ,,
Florida . .	At 6000 ,,	,,	17 ,,
	At 11,000 ,,	,,	24 ,,

Data on the relation between width of up-draught and of cumulonimbus cloud obtained from radar observations by Jones,[64] indicate that the horizontal width of the up-draught is slightly more than half the cloud width. Not all of the cloud is due to up-currents, and peripheral parts of a cumulonimbus could be due to the shear in spreading.

If the mean up-current in cumulonimbus is taken as 22 ft. per sec. (15 m.p.h.), then the amount of cumulonimbus (expressed in eighths) in convergence zones of various sizes should be not far from the following values:

$x =$	100	500	1000 mls.
$y =$ 10 mls.	3/8	3/8	3/8
50 ,,	> 1/8	> 1/8	> 1/8

Taking conditions which should produce greater cloud density, suppose that $u_1 = u_2 = 15$ m.p.h., and that $U = 0$ (i.e. when all the air converging is relieved by up-currents). Then average cloud amounts for various values of x and y would be not greater than the following:

$x =$	100	500	1000 mls.
$y =$ 5 mls.	8/8	8/8	8/8
10 ,,	4/8	4/8	4/8
50 ,,	1/8	1/8	1/8

The degree of convergence in this example would rarely be exceeded near the Equator. Cumulonimbus columns vary greatly in diameter, their average being about one to two miles with a probable maximum of five. Under conditions of maximum convergence into a zone five miles or less wide, the intensity of cumulonimbus may be 8/8, but with zones fifty miles wide cumulonimbus would be very scattered.

Consider the case when the horizontal outflow from the end of the zone is great, i.e. U = 20, with $u_1 = u_2 = 10$ and $\theta_1 = \theta_2 = 45°$. For different values of x and y the air contained in vertical up-currents penetrating the 10,000-ft. level is:

$x =$	100	300	500	1000 mls.
$y =$ 50 mls.	1475	6875	12,275	25,775 cu. m.p.h.
100 ,,	250	5650	11,050	24,550 ,, ,,
300 ,,	−4650	750	6150	19,650 ,, ,,

From this it will be seen that, for the given velocities in the streams, a cloud-free section of the boundary about 300 miles long could occur on the up-stream side of a belt of fresh winds (20 m.p.h.) 300 miles wide. For short sections of the line, divergence may occur over a critical width, causing subsidence and a cloud-free zone.

These calculations exclude the entry of down-draughts into the zone. Subsiding currents occur near convectional cumuli of quiet conditions, but it is unlikely that any great down-flow could exist for zones of convergence of the scales in these examples. Observations show that considerable stratiform cloud is usually associated with boundaries, which is evidence that there cannot be much descending air. It seems justifiable to assume that nearly all the air flowing horizontally into the zone is discharged upwards, probably entering the general upper-level flow to middle latitudes.

Towering cumulus and cumulonimbus on a boundary do not develop uniformly but are balanced with wind strength—winds are generally stronger down-stream from a cloud-free part of the boundary. This condition is frequently found from June to September, when the Southwest Monsoon covers tropical Southeast Asia north of 5° N. South of the boundary then are the Southwest Pacific Trades, which on approaching Borneo turn from southeasterlies to southwesterlies. If a typhoon forms near the Philippines at this period, strong southwesterlies to feed it develop in a wide zone along the boundary. Over both Borneo and Malaya the boundary is

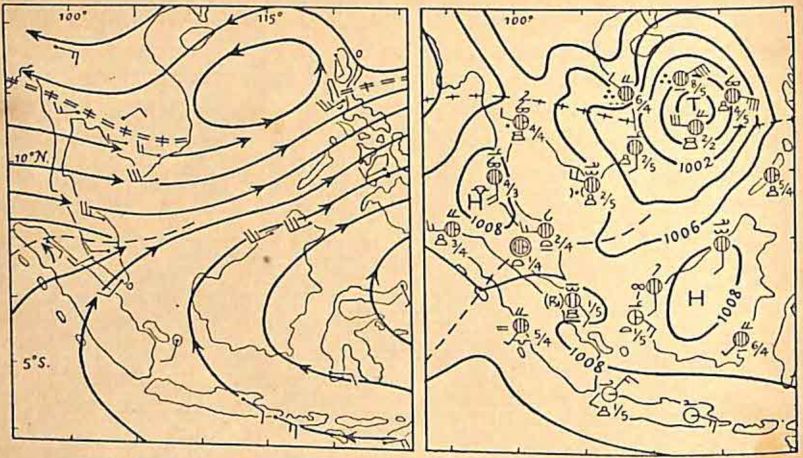


FIG. 60 Strong Outflow along a Boundary

(*a*) Winds at 5,000 ft. 0000–0300 G.M.T., 19.9.49

(*b*) Synoptic Chart, 0600 G.M.T., 19.9.49

indistinguishable, partly because the streams are nearly parallel, partly because subsidence takes place along the line west of the strong outflow (Fig. 60).

4. Air-stream Patterns and Equatorial Convergence Lines

Three types of equatorial air-stream boundary may be recognised, those sections of it developing cumulonimbus being known as 'Equatorial Convergence Lines'.

The types are:

(*a*) A Simple or Stable Air-stream Boundary, related to parallel non-convergent streams (Fig. 61); it is free from towering cloud, although altostratus may result from shear if the air velocities differ greatly.

(*b*) A Convergence Line, occurring when two streams are emphatically convergent (Fig. 62). In normal or strong convergence cloud formations will vary from 8/8 cumulonimbus for a zone up to 5 miles wide to below 1/8 for a zone 50 miles wide. The total cloud cover may be greatly increased if stratocumulus and small cumulus are present, and horizontal shear may induce altostratus formations far back into one or both streams.

(*c*) An Inactive Air-stream Boundary, associated with strong outflow down stream and marked by a comparatively cloud-free zone (Fig. 63).

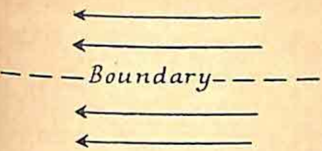


FIG. 61 Simple Air-stream Boundary

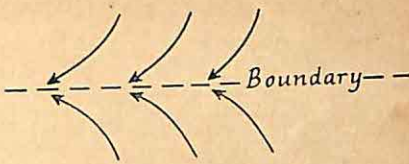


FIG. 62 Convergence Line

5. Convergence Zones near Air-stream Boundaries

A strong boundary out-flow (Type (*c*) above) leads to more complex patterns. In cases involving a typhoon, for example, the strong stream along the boundary may be continuous and non-converging for hundreds of miles. A velocity convergence may be built up at times along a line transverse to the boundary: air streams A and B (Fig. 64) may be of similar origin and properties, but may have diverged slightly so that the portion of stream A entering zone XXX may contain high outflow velocities near a relatively cloud-free section of the boundary CD. Convergence then exists over zone XXX, where cumulonimbus develops in scattered and non-linear forms. Owing to the common source, zone XXX is not a true air-stream boundary.

The 'Convergence Zone' (XXX) may move down-stream, but usually remains stationary because its existence depends on a distant feature (probably hills) which initiated the division of the stream. It is often associated with a weak trough: at higher levels the stream-line discontinuity becomes less marked until at a certain altitude one stream will be found passing completely across the zone. Fig. 65 shows the stream at 10,000 ft. above the region in Fig. 64. Convergence zones within a stream are temporary and customarily obliterated after a few days when the streams merge again nearer their source.

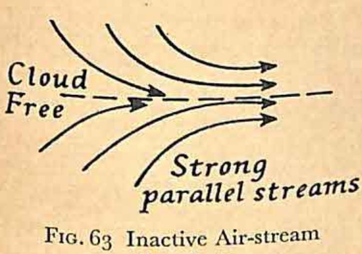


FIG. 63 Inactive Air-stream Boundary

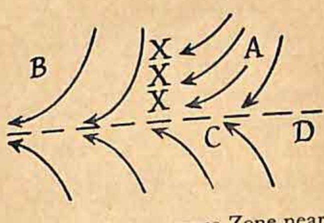


FIG. 64 Convergence Zone near Boundary

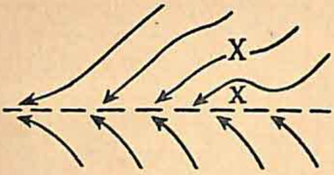


FIG. 65 Flow above Convergence Zone

6. Clouds at Sloping Boundaries

The majority of boundaries are sloping rather than vertical as in the above examples. Mather[60] established for sloping boundaries a correlation between rainfall and the position of the boundary at 3000 ft. The strongest factors initiating those upward currents which relieve horizontal convergence are ground-heating and orographic uplift, with the result that greatest cloud development should occur at the ground position of the boundary. A correlation closer than Mather established should exist between the place of maximum rainfall and the location of the boundary below 3000 ft.

Boundaries of very gentle slope have only slight activity (if any), mainly near the intersection with the ground, indicating that air converging at great heights does not necessarily initiate up-currents. Cumulonimbus even over the ground position of a gently sloping boundary is sparse, because only a shallow layer contributes to the development and because convergence at upper levels is too remote to assist developing the up-draughts.

The three classes of cloud structure on a boundary may be distinguished:

(1) When slope is almost vertical, converging air at all levels may contribute to cloud growth over the ground line and, given adequate convergence, the boundary is well-marked, comparatively narrow and intense (Fig. 66).

(2) When slope is very gentle (approaching horizontal), cloud is scattered and mainly near the ground boundary, either on it or on the same side of it as the sloping surface (Fig. 67). Cumuliform may be sparsely scattered over a zone hundreds of miles wide, a marked line with concentrated cumulonimbus being rare. Cumulo-

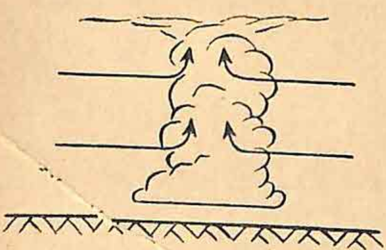


FIG. 66 Cloud on Vertical Boundary

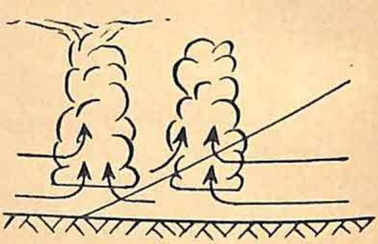


FIG. 67 Cloud on Sloping Boundary

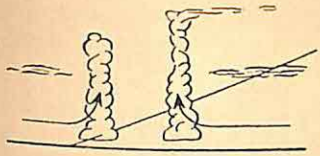


FIG. 68 Altostratus about the Boundary

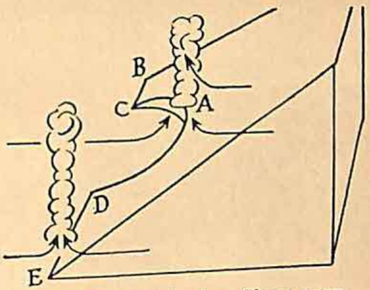


FIG. 69 Cloud when Slope near Vertical

nimbus does not often grow at great distances up the slope because convergence normally diminishes at higher levels. Analysing the slope of many boundaries over Southeast Asia (Chapter XI), we find that it is always changing, and gentle slopes are temporary because at higher levels they are nearly always undergoing rapid displacement by the winds in one of the streams. In such a case, convergence is negligible, and there is little to induce cumulonimbus to develop away from the ground boundary. Shear may be great at many places in either stream, accounting for sheets of altostratus frequently found unrelated to the slope (Fig. 68).

(3) Intermediate to these classes is an appreciable and nearly vertical slope, frequently found near the Equator. In this, the degree of cloud development varies greatly, though generally less than when slope is vertical. A typical case (Fig. 69) shows cumulonimbus scattered over a zone up to 100 miles wide, or strung out raggedly along the ground boundary or slightly to its sloping side. When a cumuliform column grows at a distance (A in Fig. 69) from the ground boundary both streams tend to feed it, distorting the ground boundary (BCADE) at that point. Under these circumstances, the conception of a ground boundary is almost illusory, yet still useful because the smoothly curved low-level boundary (analysed from the winds) indicates a mean position to which the actual boundary tends to return after the collapse of the separate cumulonimbus which distort it.

7. Diurnal Variation of Cloud on Boundaries

In presenting these model boundaries, an important factor has so far been neglected—the diurnal variation of temperature. Although convergence governs cumuliform development on a boundary, ground heating and cooling greatly affect air stability in equatorial

and tropical regions. The following figures for Singapore and Kuala Lumpur show that the diurnal temperature range is much greater inland than at the coast:

Station	*Mean Max. Temp.*	*Mean Min. Temp.*	*Mean Range*
	° F.	° F.	° F.
Singapore	86·6	75·0	11·6
Kuala Lumpur	90·5	72·0	18·5

Upper-air temperatures over the region are nearly always conditionally unstable, and an increase of ground temperature by 18° F. ensures that the lowest layer is conditionally unstable if not absolutely so. Therefore, where a boundary exhibiting horizontal convergence lies across a land mass, it induces considerable cumulonimbus during the daytime. Such development exceeds the amounts previously estimated in this chapter, because low-level air is drawn into the system from offshore (sea breezes) to feed the rising currents along the boundary. This inflow from the sea decreases low-level convergence on the boundary offshore. Consequently cumuliform is rare during day-time on a boundary over coastal waters (although shear may form altostratus there), as illustrated in Fig. 70, which shows a ground boundary oriented east–west across Malaya. By day considerable cumulonimbus lies in a wide zone (AA of Fig. 70) over the land and, if the streams are unstable, clouds developing on the boundary may merge with the intense diurnal build-up of cloud on land (BB). Offshore, and particularly to leeward of the land, cloud is probably negligible on the boundary by day, owing to divergence in the streams (C, C′ of Fig. 70). Farther than 20 miles from shore[2] the sea-breeze effect is small, and the boundary is likely to be marked by cumulonimbus, towering cumulus or cumulus sufficient to discharge what is converging from the opposing streams.

At night conditions are reversed: radiation cools the land and reduces temperatures in the lower air, promoting vertical stability. Nocturnal cooling causes the onset of a land breeze around the coast-line, a low-level divergence compensated by subsidence in the now stabilised atmosphere. Consequently, a boundary over land at night is rarely marked by cumuliform, though it is not necessarily cloud-free because if an altostratus sheet had spread along the boundary during the day, it could persist there overnight.

Offshore along the boundary at night, excessive cloud is likely on an exposed coast for two reasons:

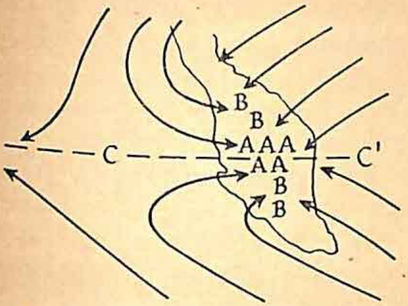


FIG. 70 Surface Flow on Boundary by Day

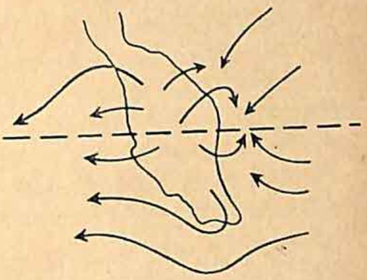


FIG. 71 Surface Flow on Boundary by Night

(1) Low-level convergence is increased by the addition of the land-breeze outflow (Fig. 71).

(2) Although diurnal variation of sea temperature is negligible, the temperature above any pre-existing cloud layer may fall by radiation, setting up a steeper lapse-rate over the sea, and promoting convection. Cloud on an offshore boundary is in consequence excessive at night.

Farther than 10 miles from the land mass, land breezes are not usually experienced,* causing the density of cumuliform to depend on the degree of convergence in the air currents. Even over the ocean, however, cumulonimbus is likely to develop more by night than by day, because radiation from the tops of pre-existing cloud layers may, by increasing the lapse-rate, promote convection.

Rough rules may thus be formulated for predicting the diurnal variation of density of cumulonimbus or towering cumulus on the boundary between two moist currents. Normals of cumuliform density for varying zone widths may be established corresponding to the degree of convergence in the streams (see page 120 for normals in cases of moderate and strong convergence). Greater densities and narrower zones are typical of vertical boundaries, whereas slighter densities and wider zones accompany boundaries of gentle slope. Normal densities are found over the seas by day and often by night. Excessive densities are over land by day and offshore by night, particularly to windward near the land-masses. Densities below normal occur over land at night and offshore by day when cumuliform leeward of the land masses may be negligible.

* The distance varies according to the topography, and a land breeze reinforced by a katabatic flow from high ranges may extend to a greater distance offshore.

8. Diurnal Change of Position of a Boundary

Diurnal temperature variation, the dominant cause of departures from normal cumuliform density, may also displace the boundary. Where it lies close to a large land mass, the boundary suffers low-level distortion, related to the land- and sea-breeze régime. This occurs[57] sometimes in the Malayan region during May and June when the boundary between the Indian Southwest Monsoon and the Southeast Trades has a northwest–southeast orientation near the west coast of Malaya.

(1) During evenings, the land breezes from both Malaya and Sumatra produce maximum low-level convergence in the Straits of Malacca, inducing more cumulonimbus than is normal. By midnight, at about 3000 ft. or higher, the Southeast Trades and the Monsoon Westerlies cross the high country of Malaya and Sumatra to align the upper air-stream boundary also along the Straits.

(2) During mornings, rising land temperatures may transfer maximum low-level convergence to the Sumatran or Malayan mainland and, as successively higher layers contribute to cumulonimbus, the boundary becomes aligned with the ranges of Sumatra or Malaya, reaching up to considerable altitudes above them.

9. Line Squalls on Moving Boundaries

Because air streams frequently flow asymptotically at their common boundary, the boundary moves slowly, and rarely in the manner of that line squall usual on temperate-latitude cold fronts. When a squall does occur (through increasing velocities within a stream), its direction changes gradually until it travels along the boundary. The velocity component athwart the boundary is small, so that only a small displacement of the boundary may be associated with a strong squall and bad weather.

A complication occurs when velocities increase in a stream containing a line squall (AA of Fig. 72) moving down the boundary. Suppose the boundary itself is regularly but slowly being displaced from XY to X′Y′, the squall AA moving down the boundary is likely to induce a secondary squall BB in the other stream, so that an observer at P may experience a squall and weather from the southwest before

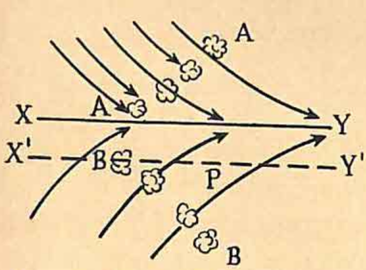


Fig. 72 Line Squalls and Induced Squalls

the arrival of the main boundary from the north. (See 'Induced Squalls' in Chapter XIII.)

10. Convergence within a Stream

Most precipitation and cumuliform cloud in equatorial regions is brought about by three factors effective singly or in combination; they are:

(i) Convergence on air-stream boundaries, (ii) orographic ascent, and (iii) ground heating.

Some rain falls in minor disturbances within a stream and remote from land masses. These disturbances are so diverse that at first it appears difficult to include them in one description. Wood[65] traced 'disturbance lines' of cumulonimbus and rain crossing Northeast Borneo, but how far they had travelled previously is not known. There is no record of a disturbance, linear or zonal, crossing the South China Sea from Indochina to Eastern Malaya or North Borneo, so that Wood's 'disturbance lines' were probably local or short-lived. Sudden periods of general heavy rain in East Malaya occasionally begin after an abrupt rise of pressure near Hong Kong, brought about less by the approach of a linear disturbance and more by relief and increased wind velocities resulting from an increased pressure gradient over the South China Sea.

'Surges' have been mentioned in connection with the Indian Southwest Monsoon. A surge is a disturbance of large extent moving with the strengthening wind and accompanied by heavy rain, increasing cloud and sometimes a slight wind-shift. Whether or not a pressure trough accompanies the surge is doubted; ground isobaric analyses at 1-millibar intervals fail to show it, so that the trough must be very shallow if it exists at all.

Riehl[54] describes a type of disturbance called an 'easterly wave', consisting of a weak pressure trough in and transverse to the easterly Trades, associated with a wind-shift and followed by increasing cloud and rain. Such a wave travels westward at about 400 miles per day. The 'easterly wave' resembles a convergence zone in many features, as also do 'surges' and 'disturbance lines'. Common to all are increased cloud and rain and a variation of wind direction or velocity across each.

These disturbances occur within a stream and thereby differ from boundary features. The various terms for them serve mainly to define locality, and all might aptly be described as 'convergence zones' where (regardless of other cloud types) cumuliform is evident, or as 'shear lines' where cumuliform is absent but altostratus exists

to imply wind-shear. It is also desirable to term them 'moving' or 'stationary'.

11. Doldrums

Before the details of equatorial wind streams were known, the equatorial region was assumed to be one of light winds without definite direction of flow and known as the 'doldrums'. Later, when isobaric and stream-line charts came into general use, smaller regions of light variable winds were marked as 'doldrums'. Theories concerning their structure involved the suggestion that showers were fairly widespread within them, and that lines of cumulonimbus or altostratus marked their edges. Increased pilot-balloon stations in Southeast Asia during recent years have shown that, although ground winds may be light and variable over large areas, a stream flow is discernible aloft; thus the extent of 'doldrums' on the upper stream-line chart diminishes with increase of knowledge. While the monsoons blow, regions of light winds aloft never occur. The term 'doldrum' may ultimately be eliminated from the meteorological vocabulary as far as upper-winds are concerned, but 'doldrums' are still shown on the charts of many services.

CHAPTER XI

Circulation over Southeast Asia

1. Analysis

The march of seasons over Equatorial Southeast Asia will now be examined from typical weather charts. An idea of the locations of the surface boundaries may be obtained from the seasonal mean stream-line charts in Chapter I, which show, for example, that in April the mean position of the discontinuity between the Trades of the North Pacific (T_{NP}) and those of the Southwest Pacific (NT_{SP} (Equat.) or NT_A) is near Southern Borneo (Fig. 5). They also show that the mean April position of the boundary separating the Equatorial Westerlies (T_{SI}) from the North Pacific Trades (T_{NP}) is oriented from north to south along Sumatra and Malaya.

It is difficult to visualise the position of the boundary by studying the mean wind streams for May, because then circulation is completely changing with the advance of the Indian Southwest Monsoon. Adequately to describe the seasonal changes involves examining charts not of mean positions but of sample seasonal phases.

For most of the year, two main boundaries exist between Indo-china and Northern Australia. They usually have separate identities in the west of that region but are combined in the east. The average structure is shown in Fig. 73. The northernmost (AB) is the 'Northern Equatorial Air-stream Boundary'; the southernmost (BC) is termed the 'Southern Equatorial Boundary'; where they coincide (BD) is called the 'Combined Boundary'.

During the northern summer, the Northern Boundary (AB) is displaced far to the north over China, where it becomes the Inter-tropical Front separating the Northeast Trades of the Northwest Pacific from the west-southwesterlies of the Indian Southwest Monsoon. At this season, the Southern Boundary (CB) separates the Southeast Trades of the Southwest Pacific from the Southwest Monsoon. Because the monsoon develops from a diversion of these Trades, the Southern Boundary may not extend far to the west.

Between the two monsoon seasons, the boundaries take up positions near the Equator. The Combined Boundary (BD) separates the Trades of the two hemispheres, frequently extending far to the west. At other times, the Northern and Southern Boundaries may be wide apart; they are separated by a belt of Equatorial Westerlies,

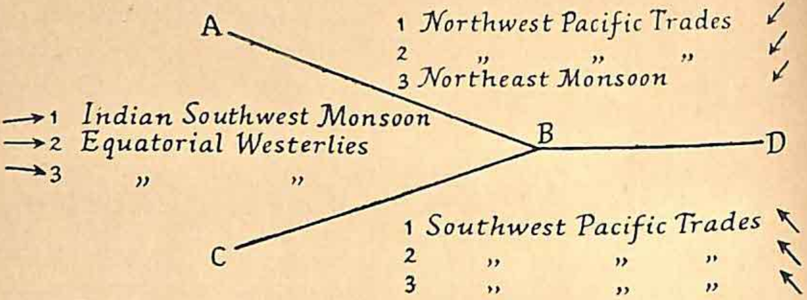


FIG. 73 Major Equatorial Air-stream Boundaries of Southeast Asia

the origin of which is sometimes uncertain during the inter-monsoon period, though charts of mean flow indicate that they frequently extend back at least to the Arabian Sea.

In the southern summer, the entire system moves southward and the line formed by the Southern and Combined Boundaries (CBD) becomes the Intertropical Front, separating streams from two hemispheres. Southwest Pacific Trades (southeasterlies) still blow south of CBD, and the Northeast Monsoon has developed north of ABD. The Northern Boundary then appears intermittently: Equatorial Westerlies may still occur south of it, composed of Northeast Monsoon air diverted to a westerly orientation over the Indian Ocean. The Northeast Monsoon may even blow directly into the Southern Hemisphere without this diversion, on which occasions the Northern Boundary is non-existent.

This system is idealised. Separate Northern and Southern Boundaries depend on the orientation of the several streams, so that there are many occasions each year when only the Combined Boundary exists. This is more evident if we examine the seasonal movements in greater detail.

2. The Equatorial Westerlies

Fletcher[66] explained the Equatorial Westerlies as a semi-permanent feature of the planetary circulation.* He stated that, if cumuliform develops markedly along the two major boundaries, probably considerable radiative cooling is occurring from the cloud tops. The region along the Equator and between the two Boundaries

* *Semi-permanent* Equatorial Westerlies are confined to the Indian Ocean, Indonesia, the Eastern Pacific and Central America. Any Equatorial Westerlies occurring elsewhere (according to C. E. Palmer in *Journal of Meteorology*, 1952, vol. 9, p. 377) are due to individual cyclonic circulations, so that it is difficult to maintain continuity in tracing the air-stream boundaries between them and the Trades.

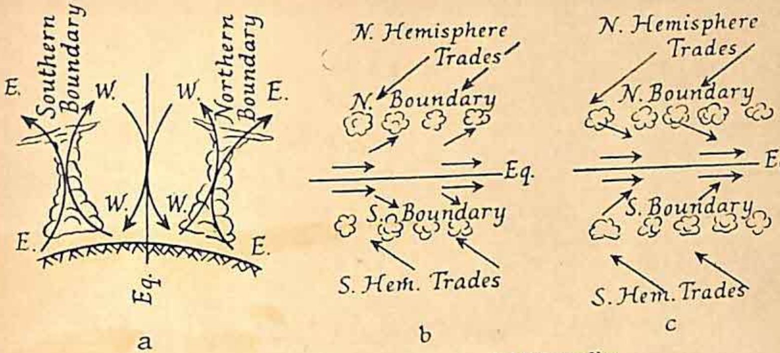


FIG. 74 Origin of the Equatorial Westerlies

then becomes a source not of warm air but of cool subsiding air (Fig. 74*a*). A separate cyclonic shear would be established on each Boundary between the cooled source and the warmer Trades of each hemisphere, and the air between the two boundaries would develop into a westerly stream. At low levels the westerlies would have a poleward component in each hemisphere flowing in to feed the up-currents along the boundaries (Fig. 74*b*), and at high levels they would have an equatorward component as the cooled air subsides from the cloud tops into the zone (Fig. 74*c*).

No data being available regarding the temperature drop possible from radiational cooling of the cloud tops, it is difficult to assess the reliability of this theory. The air rising in the cumulus clouds should cool at the saturated adiabatic lapse-rate, and that subsiding between the boundaries should warm at the dry adiabatic rate (which is greater), causing a net gain of temperature which radiative cooling must offset before the zone could be considered a source of cold air.

3. Seasonal Movements of Boundaries at Low Levels

The march of the boundary system is as follows. In July the Northern Equatorial Air-stream Boundary does not appear in Southeast Asia, having preceded the Southwest Monsoon north of 15° N. and become the Intertropical Front of China. The Southern Equatorial Boundary at this time lies in a northeast–southwest direction in the neighbourhood of Malaya (Fig. 75), separating the Indian Southwest Monsoon (T_{SI}) from a southeasterly current of Australian or Southwest Pacific origin (N_{TA}, NT_{SP} or NT_{SP} (Equat.)). Occasionally a separate anticyclonic cell lies over Borneo, but the boundary (Fig. 76) dividing it from the south-easterlies is only temporary.

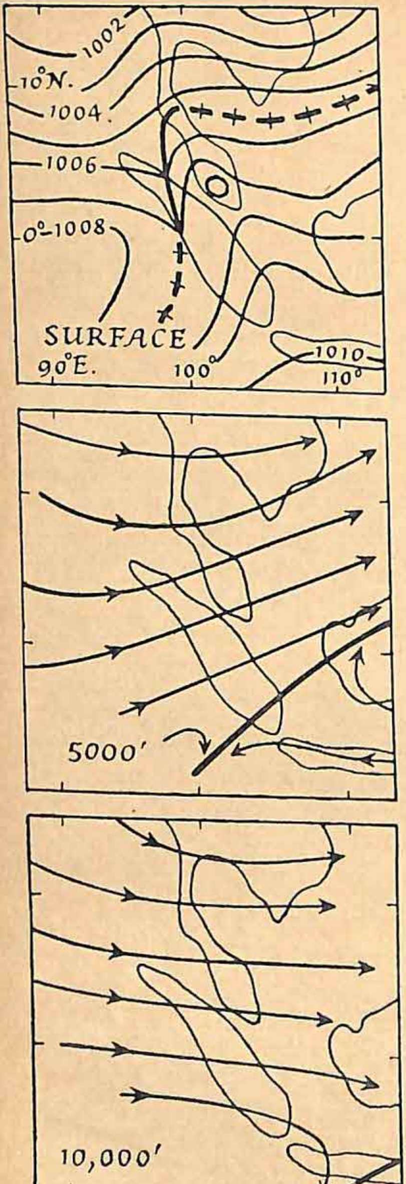


FIG. 75 Weather Chart and Streamlines, 0000 G.M.T., 5·7·48

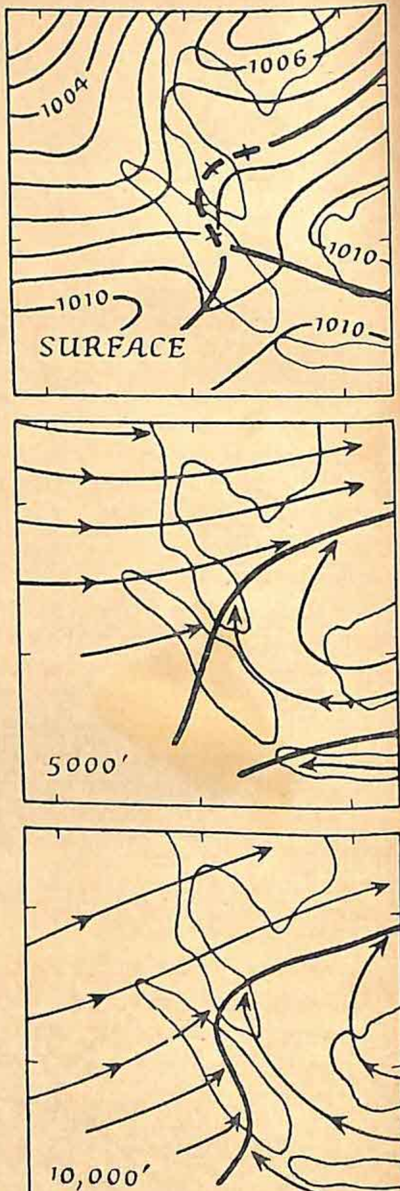


FIG. 76 Weather Chart and Streamlines, 0000 G.M.T., 8·7·48

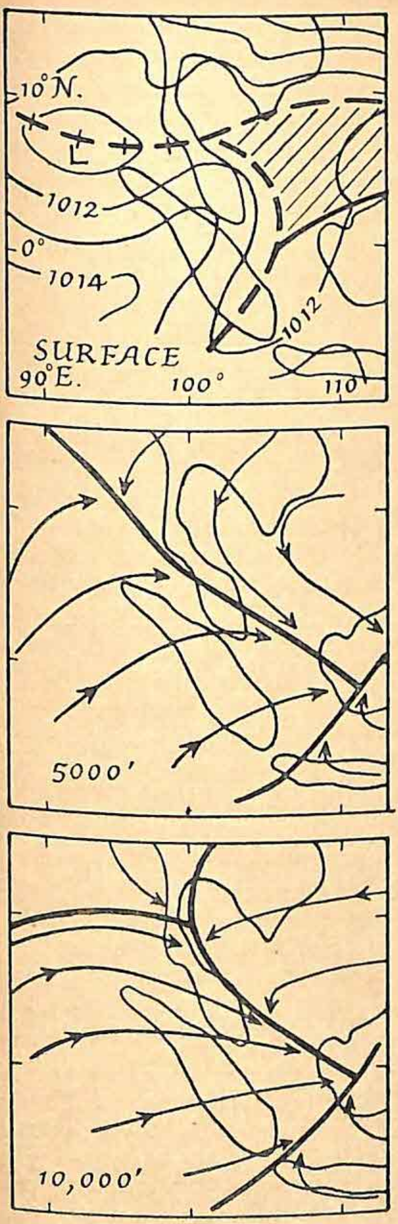


FIG. 77 Weather Chart and Streamlines, 0000 G.M.T., 27.10.47

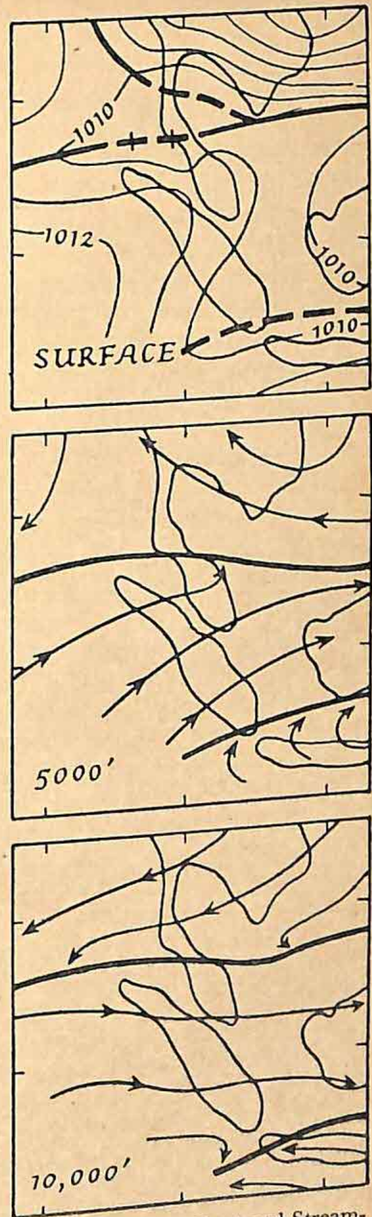


FIG. 78 Weather Chart and Streamlines, 0000 G.M.T., 10.11.47

In July, the Southern Equatorial Boundary moves between Palembang–Bangka–North Borneo in the south to Sabang–Phuket–Gulf of Siam in the north. Movements of the line are of two types: when it is oriented north–south in the Straits of Malacca, there often appears to be a diurnal fluctuation of position, displacement to the east taking place during the early morning and a slow return to the original position during the late morning. Larger displacements within the above limits occur under outside influences. Movements to the northwest are associated with slightly increased low-level velocities in the southeasterlies, probably dependent on the migratory anticyclones of the Southern Hemisphere. Others may be associated with a rapid rise of pressure over Borneo. Large displacements to the southeast occasionally follow a surge (moving convergence zone) of the Southwest Monsoon at times when the southeasterlies are weak.

August produces little change except that the monsoon surge is more frequent. This disturbance may be traced, in its four-day passage across the southern Bay of Bengal, by wind-shear, by cumuliform, and by a thickening of the altostratus or by the formation of altostratus where cirrostratus alone previously existed. The impact of the surge is obvious on the Malayan coast north of lat. 7° N., but the high mountains of Sumatra prevent the cloud increases penetrating to Malaya and eastern Sumatra. The reorientation of the monsoon from southwesterly to westerly following a surge favours the diversion into the Straits of Malacca and across Malaya of a northwesterly current up to 8000 ft. deep (see Chapter XII). Not only does this displace the Southern Boundary southeastward, but reinforced convergence promotes activity on it. Eastward movements near Malaya are invariably associated with great activity, but westward displacements cause considerably less.

During September, the Northern Equatorial Boundary continues mainly north of 15° N. (where it is known as the Intertropical Front), and the incidence of a separate Borneo high with consequently a new boundary over South Borneo is still intermittent. The Southern Boundary at this time is located through Palembang, and stretches thence to the northeast. Its movement is influenced by the diurnal oscillation and by isolated monsoon surges, and probably moving convergence zones in the southeasterlies affect the position and intensity of the boundary.

By mid-October, the Northern Boundary passes south of 10° N. Its structure is normally a single line, though perhaps occasionally widening to a zone (Fig. 77)—a situation rarely sufficiently long-lived to promote instability throughout the doldrum. In the west

also this boundary moves southward during October: under the influence of Indian northerlies, it crosses the Bay of Bengal to Sabang and south of Ceylon.

In October also, the Southern Equatorial Boundary is oriented northeast–southwest, mainly through Bangka and northwest of Borneo where it joins the Northern Boundary. The former separates Southwest Pacific southeasterlies or easterlies (NT_{SP} or NT_{SP} (Equat.)) from Indian Ocean westerlies (T_{SI}), and the Northern and Combined Boundaries separate both of these from the northerlies (NP_s) or northeasterlies (T_{NP}) of the Northern Hemisphere.

By the end of November, the Southern Boundary lies east-northeast to west-northwest near Java, and the Northern Boundary moves to Malaya with the onset of the Northeast Monsoon (Fig. 78). Isolated weak depressions may form and fill east of Malaya. The origin of the westerlies between the two boundaries is now doubtful, and the Northern Boundary can no longer be considered a true Intertropical Front.

In early December, further advance brings the Northern Boundary to the Equator whence it may still return to Southern Malaya. The Southern Boundary is crossing Java, and a well-defined belt of Equatorial Westerlies separates the boundaries (Fig. 79). Each of the two boundaries varies widely in intensity at this stage, alternately dying and intensifying following slight variations in the orientation of the westerlies.

During January and February, the Southern Boundary lies between 10° and 12° S. (Fig. 80), retreating a little to the north when it is reinforced by weakening Southern Hemisphere cold fronts (which may be traced by changes at Cocos Island). The Northern Boundary is intermittent with a mean position through Bangka and south of Borneo, but no longer distinct west of Sabang. West of Sumatra the Northeast Monsoon is diverted into Equatorial Westerlies, which at times disappear completely when the Northern Boundary travels south, allowing a northeasterly flow at all levels to Java. The Southern Boundary might now be termed a true Intertropical Front and, when the two boundaries merge, their intensity is usually pronounced despite the easterly flow on both sides. The Northern Boundary is now an uncertain division between true northeasterlies and those which have turned by crossing the Equator; it frequently disappears to form again a few days later some distance away.

By March, the origin of the Westerlies becomes complex once more, and the Southern Boundary may no longer be considered a true Intertropical Front. Southwest Pacific southeasterlies advance

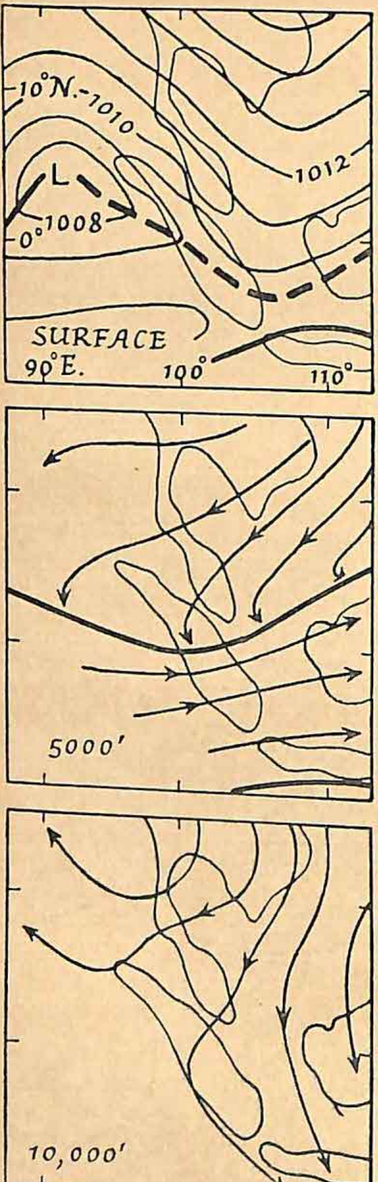


FIG. 79 Weather Chart and Streamlines, 0000 G.M.T., 28.12.47

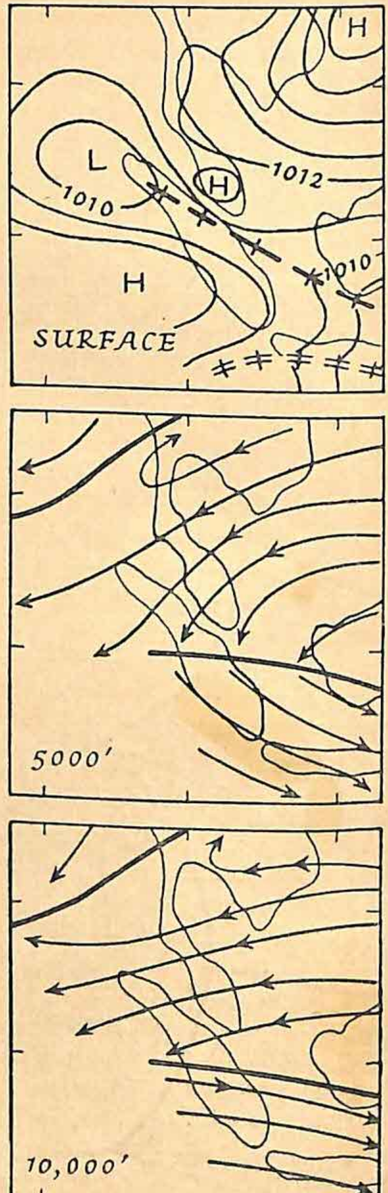


FIG. 80 Weather Chart and Streamlines, 0000 G.M.T., 22.2.48

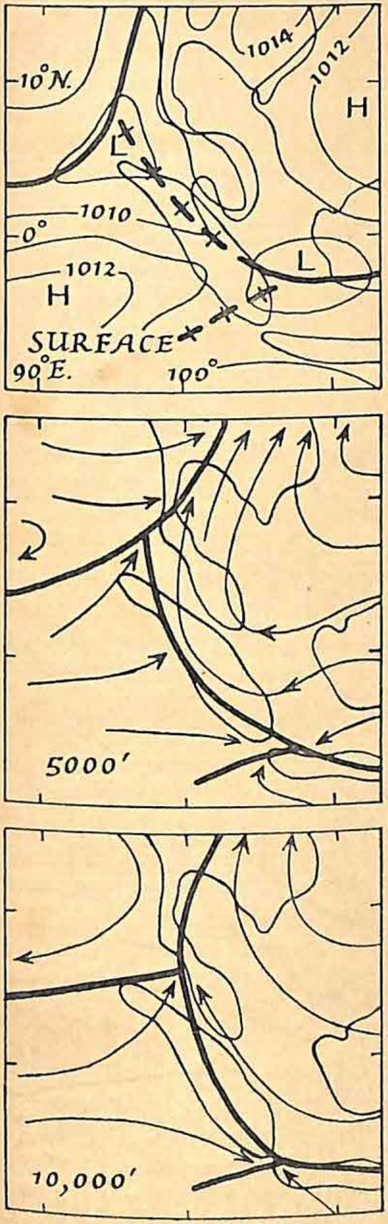


FIG. 81 Weather Chart and Streamlines, 0000 G.M.T., 20.4.48

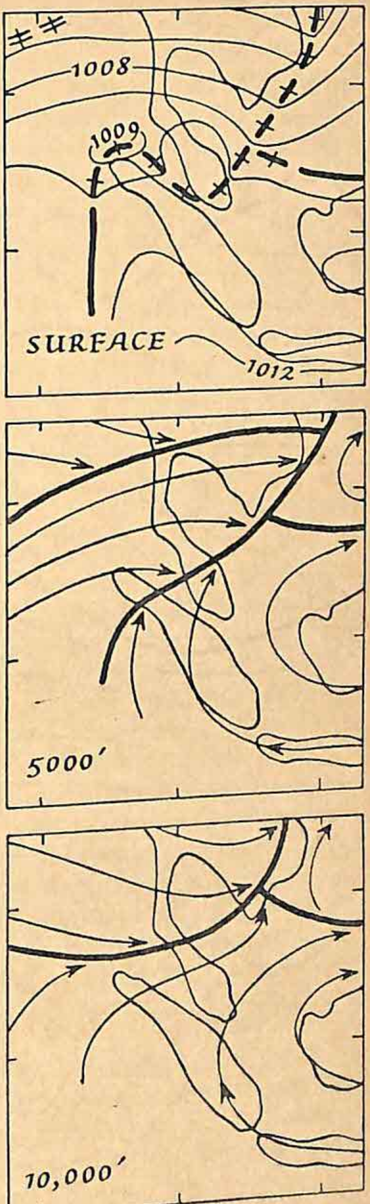


FIG. 82 Weather Chart and Streamlines, 0000 G.M.T., 28.5.48

the Southern Boundary to the north of Java, while both the Equatorial Westerlies and the Northern Boundary may vanish.

Further advance in April combines the two boundaries east of 106° E. as a single line (Fig. 81) on which convergence is frequently intense. West of 106° E. two Equatorial Boundaries still exist—the Northern through the Straits of Malacca and across Sabang to Ceylon, and the Southern to the southwest through the Soenda Straits. Westerlies from both hemispheres still enter between the boundaries and may reach South Borneo. An additional north–south boundary in the eastern Bay of Bengal, which separates dry northerlies (probably NT_1) from an equally dry portion of the Northeast Monsoon, may now exhibit slight activity when the Northeast Monsoon is diverted to southerlies (Fig. 81).

The northward march continues in May, when the Combined Boundary moves across Borneo. For a while, the Northern Boundary fluctuates across North Sumatra and Malaya. The Equatorial Westerlies gradually broaden and strengthen, deviating to a more southwesterly track until by the end of May (Fig. 82) they cover the southern half of the Bay of Bengal, Malaya and Thailand. Marking the onset of the Southwest Monsoon, the Northern Equatorial Boundary advances to Ceylon, Southern Burma and Indochina as the Intertropical Front. A weak, though increasingly active, Southern Boundary has now advanced over Malaya to separate the monsoon from the Southwest Pacific southeasterlies.

During June, continued development of the southwesterlies, now clearly crossing the Equator from the Southern Hemisphere, forces the Northern Boundary (as Intertropical Front) from 12° N. to 20° N. The Southern Boundary, now fairly fixed, varies in intensity: it occasionally comes southward to Singapore from its mean position, which is from Palembang, along the Sumatran ranges or Malacca Straits to Penang, and thence across Borneo (Fig. 83). Few disturbances in the Trades or surges in the monsoon penetrate the region at this month.

4. Displacement of Boundaries at Higher Levels

The seasonal movements of the Northern and Southern Boundaries at upper levels resemble those at the ground (Figs. 75—83). Observations of the slope of the boundary surfaces have shown[57] that it may be either side of the vertical. The average slope of the Northern Equatorial Boundary (computed from wind discontinuity at various levels) is 1 in 100 from the ground to 5000 ft., 1 in 75 from 5000 to 10,000 ft., and 1 in 70 from the ground to 10,000 ft.

The slope of a boundary surface is not uniform with height nor

uniform along different parts of the same boundary within any one layer of the atmosphere; one section of the slope may be on one side of the vertical and another section on the opposite side. That the slope of the Northern Boundary from ground to 10,000 ft. is steeper than the slope from ground to 5000 ft. (and steeper than that from 5000 ft. to 10,000 ft.) suggests that there are many occasions when the sense of the slope changes with altitude.

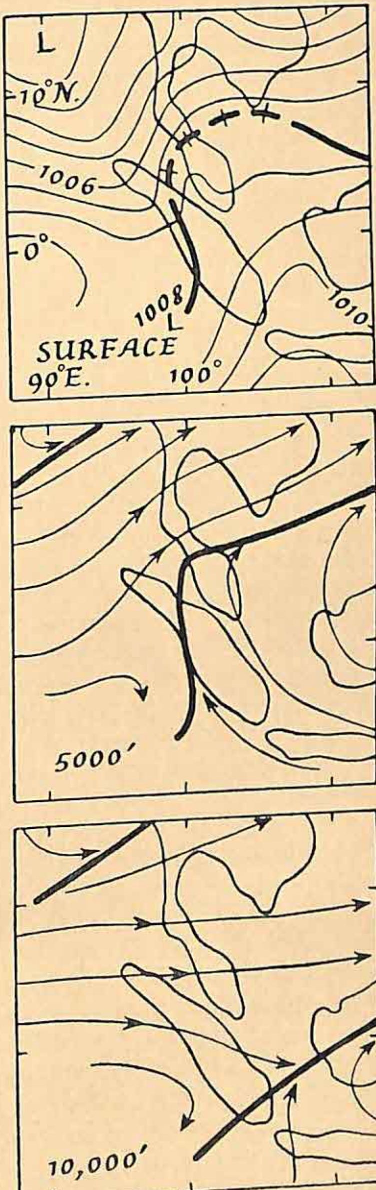


FIG. 83 Weather Chart and Streamlines, 0000 G.M.T., 9.6.48

The slope of the Southern Equatorial Boundary is generally less than that of the Northern: its mean slope from the ground to 5000 ft. is 1 in 220; from 5000 to 10,000 ft. it is 1 in 240; and from the ground to 10,000 ft. it is 1 in 210. Once again the slope in the whole layer up to 10,000 ft. is steeper than that in either of the component layers, so that we must assume the sense of slope of the Southern Boundary to be sometimes reversed with height.

The slope of the Southern Boundary varies from day to day and cannot be described adequately by a mean. Most frequently the surface slopes upwards to the east, southeast or south, with the Equatorial or Monsoon westerlies overlying the Southwest Pacific easterlies or southeasterlies. The slope to the southeast is usually gentle, and it approaches the horizontal on occasions when upper wester-

lies sweep across the equatorial zone as far as Northern Australia. This extreme case usually occurs when the westerlies are reinforced temporarily or replaced by a current of different origin—such as an advance of diverted Northeast Monsoon air at upper levels. At other times the slope increases to approach the vertical, or even changes sense so that the gradient is upwards in a northerly or northwesterly direction.

The mean period of recurrence of minimum slope of the Southern Boundary[57] is about 9½ days; the shortest observed period has been 7 days and the longest 12 days. The mean minimum slope to the southeast is approximately 1 in 360, and the mean slope in the maximum (or near vertical phase) is 1 in 20 to the northwest. Accepting these means, the slope of any section of the Southern Equatorial Boundary at any particular time may be roughly predicted by considering the period elapsed since a time when slope was vertical or a minimum to the northwest.

The periodic changes of slope of the boundary surface may follow slight fluctuations of density as well as of wind speed in the southeasterlies of the Southwest Pacific Trades, probably dependent on the relative positions of the migratory anticyclones of higher latitudes in the Southern Hemisphere. When a major anticyclone is located over Western Australia, the southeasterlies follow a short track over Australia from the Great Australian Bight or the western Tasman Sea. If the anticyclone travels eastward, part of the southeasterly current pursues an easterly or northeasterly course. By the time its centre has reached New South Wales, this air, following a long tropical sea-track, may contain temperatures slightly higher than those of the Indian Monsoon.

The Northern Boundary also may change slope in similar periods, but its slopes are much nearer the vertical and their rhythm more difficult to detect.

5. Short-period Displacements of Boundaries

Displacements of the major boundaries at very high levels, apart from the seasonal changes, are less than those at middle levels. Surface movements are frequently accompanied by displacements at 10,000 ft. but none at 20,000 ft. These differences are probably brought about by topography, examples being common in the Malayan area during June and July. During this period the ground-level Southern Boundary lies northeast–southwest across Malaya and Sumatra, separating the west-southwesterlies of the Indian Monsoon from the southeasterlies of the Southern Trades.

If these Trades should advance at the ground because of increased

velocities or reorientation, their progress to the west is checked by the mountains of Sumatra, and the ground boundary may be forced far north over Malaya and the Straits of Malacca because the southeasterly flow is diverted to southerly by the ranges. Thus the ground-level boundary may be deformed to lie along the Sumatran ranges and thence eastward from Sabang. Relief effects do not greatly influence conditions at 10,000 ft., and the boundary at that level, although perhaps advancing, will probably not suffer deformation (Fig. 75).

Relief may also influence southeastward movements of the Southern Boundary. Suppose that a boundary initially lies along the Sumatran ranges and stretches eastward from Sabang (as in Fig. 75). If it should be displaced eastward by the Indian Monsoon, the advance at 10,000 ft. may be substantial. At low levels the ranges may retard boundary movements until a surface current of air is diverted down through the Straits of Malacca as a northwesterly wind. From these considerations it will be seen why the slope of a boundary surface may change considerably over a period of a few days when near high ground.

6. Flow at Very High Levels

During the inter-monsoon periods (March to mid-May and September to mid-November), the mean wind at 20,000 ft. over Equatorial Southeast Asia is from an easterly direction at an average speed of 5 knots.* During the period June to August, the mean speed of the easterlies along the Equator is 10 knots and, owing to the incidence of monsoon southwesterlies, the average wind over Tenasserim is light southwesterly (Fig. 84).

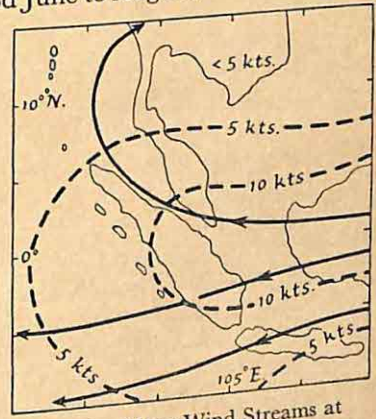


Fig. 84 Mean Wind Streams at 20,000 ft., June–August

From December to February, winds at 20,000 ft. are generally light easterlies at low latitudes, but the Northeast Monsoon has influence over the South China Sea where the mean wind is from the northeast and light (Fig. 85). The wind is southwesterly over Tenasserim, due to the occa-

* Upper-wind speeds are always in knots; surface winds usually in m.p.h.: 1 knot = 1·15 m.p.h.

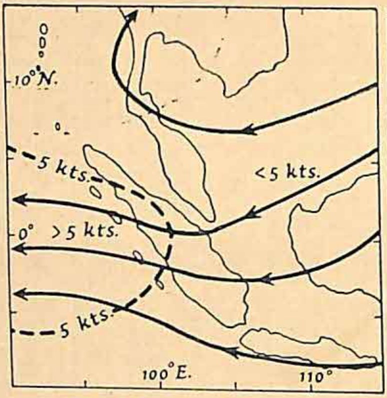


FIG. 85 Mean Wind Streams at 20,000 ft., December–February

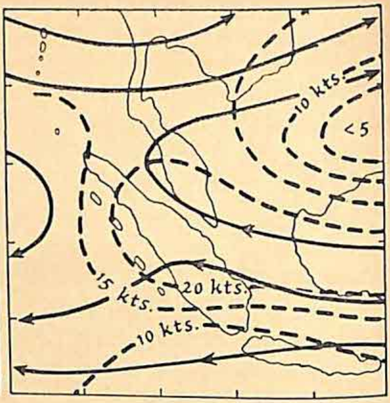


FIG. 86 Mean Wind Streams at 30,000 ft., December–February

sional lowering of upper westerlies which are well developed then.

At 30,000 ft. the pattern is simpler and, except during the period December to February, the mean wind is from nearly east. The average speed of these easterlies is 5 to 10 knots from March to May, 15 knots from September to November, and 20 to 25 knots from June to August. Indian Monsoon southwesterlies at this level are negligible.

The 30,000-ft. pattern for the December–February season is different from that of the rest of the year. The easterlies from December to February average 20 knots along the Equator (Fig. 86), while over Tenasserim and Thailand there is a constant belt of strong westerlies.

Above 30,000 ft., northeasterly to easterly directions are general over most of the area, except during December to February when the general orientation is slightly south of east. The mean speed is about 15 knots from March to May, and during the remainder of the year speeds vary between 30 and 60 knots.

Easterly directions and decreasing speeds continue up to the base of the stratosphere—between 54,000 and 66,000 ft. Within the stratosphere there is an abrupt change to westerly or southwesterly winds which, slowly increasing with height, attain about 20 knots at 70,000 ft., above which little is known about the wind structure over the Equator.

CHAPTER XII

Examples of Air-stream Analysis

In this chapter will be examined the synoptic charts and upper-wind charts of Figs. 87–94 to demonstrate how air streams are analysed from them.

On 29th July 1949, the Indian Southwest Monsoon covered the Bay of Bengal, Thailand and Indochina, while the Northern Equatorial Air-stream Boundary* was lying along lat. 20° N. in the China Sea. The Southern Equatorial Air-stream Boundary—the southern fringe of the monsoon—lay inactively across northern Sumatra while the Southeast Trades covered the Indies.

On the following day the Southern Boundary moved southward over Malaya. Rapidly for these latitudes, the boundary was advancing at 20 m.p.h. on the west coast and at 15 m.p.h. on the east, accompanied by squalls and thunderstorms and by changes of wind at most levels. The weather charts showed little ambiguity and conveniently illustrate the method of analysis.

1. Situation at 0000 G.M.T. (0730 L.T.) on 29th July 1949

In analysing this situation we are primarily concerned with deciding the location of the Southern Equatorial Air-stream Boundary. Let us first consider the stream-lines at 10,000 ft. (Fig. 87), which is practically free from orographic influences. The boundary lies southeast of Singapore because wind observations show that the monsoon westerlies cover all Malaya and northern Sumatra. The Trades are northeasterly over Northern Java, and an observation at Billiton, together with an aircraft report, indicates that the Trades extend close to Singapore. We must treat both these wind observations with reserve because they were not made at the correct observing time, the one being at 0800 hours on the 29th July and the other at 0600 hours.

Northeast of Singapore, an aircraft has reported northeasterlies. Though it is difficult to decide which air stream contained this wind, the direction is so different from the main flow in both streams that the observation was probably in that region of variable winds characteristic near the boundary. The most probable location for

* The symbols for representing the boundaries on charts are in Appendix B.

the boundary at 10,000 ft. is shown in Fig. 87, though possibly lying farther to the southeast.

Wind observations at 15,000 ft. are too scattered to throw further light on the matter (Fig. 87). The north-northwesterly at Kuantan indicates that the boundary is south of that point, justifying us in deciding that this observation is in the slowly deviating stream over Malaya, and that the northeasterly at Singapore is also in the monsoon. This is speculative: the boundary could lie anywhere south of Kuantan.

At 7000 ft. (Fig. 87) the observations are sufficiently dense to define the position of the boundary. The monsoon stream, diverted to northwesterly, obviously extends as far south as Bangka. It is north of Djakarta and probably north of Billiton, but the observation plotted for Billiton may not be relevant because it was made hours before the others. Since the boundary has now been established as southeast of Bangka at 7000 ft., there is the possibility that at higher levels also it lies farther southeast than it is shown.

Wind changes with height at Labuan give an indication of the upper-level boundary over North Borneo. At 7000 ft. the monsoon stream is established there on the evidence of the northwesterlies. At 5000 ft. and lower the wind is from the northeast, and this change within 2000 ft. is so great that, although the northeasterlies might be diverted monsoon air, they are more probably Trades.

If we draw the 5000-ft. stream-lines (Fig. 87) to include those Labuan northeasterlies in the Trades, it appears that the stream should be flowing in an arc over the South China Sea and western Borneo, a conclusion confirmed by the presence of northwesterlies over southern Borneo. A minor air-stream boundary located over the Java Sea separates the true Trades from a diverted part of them.

Examining the situation between 5000 and 1000 ft. (Fig. 87), we deduce that the Trades, either as southerlies or southeasterlies, cover most of Malaya, because air flow there conforms with the influence of orography and the Sumatran ranges are hindering the entry of the monsoon. The boundary lies north of Penang, though it is difficult to determine the precise location near Kota Bharu whose west-southwesterly wind might be in either of the streams. The direction of the wind at 1000 ft. at Kota Bharu indicates it may be a land breeze and not related to the true stream-flow aloft.

Having established where the air-stream boundaries are at 1000 ft., let us project these positions on the surface chart (Fig. 88). At 1000 ft. the Southern Equatorial Air-stream Boundary lies north

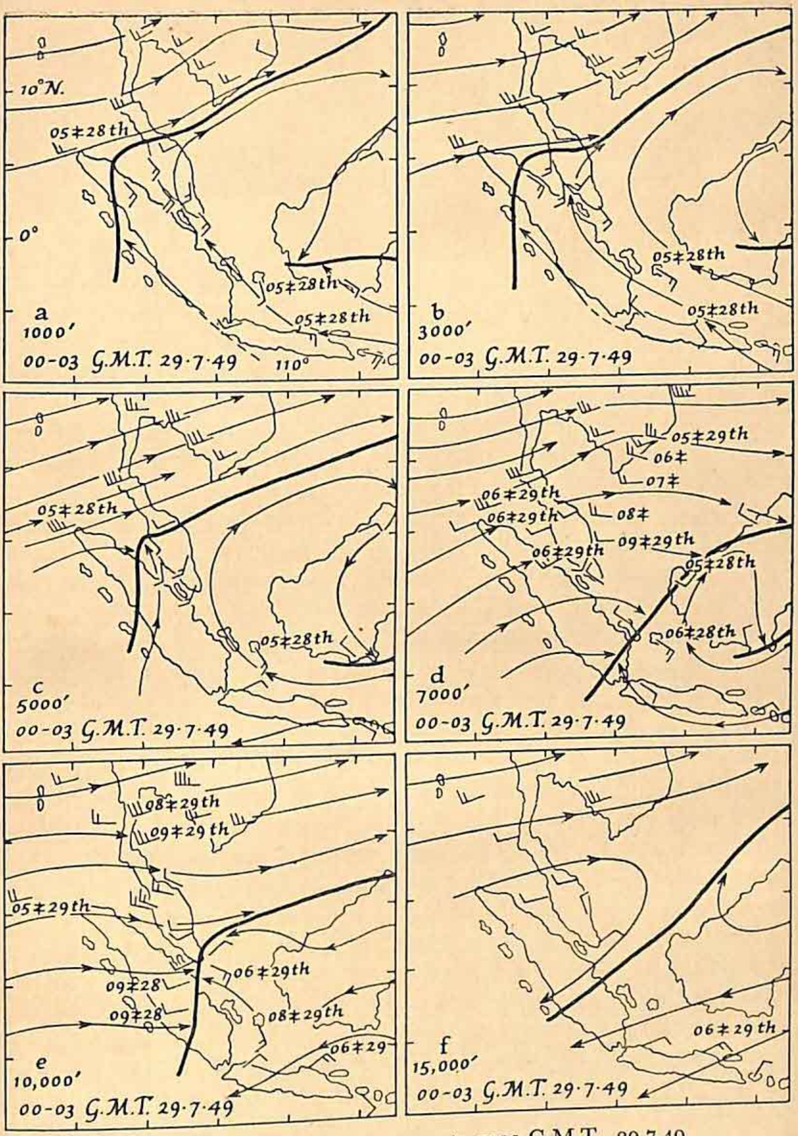


FIG. 87 Upper-wind Charts, 0000 to 0300 G.M.T., 29.7.49
(7.30 to 10.30 a.m. Malayan time)

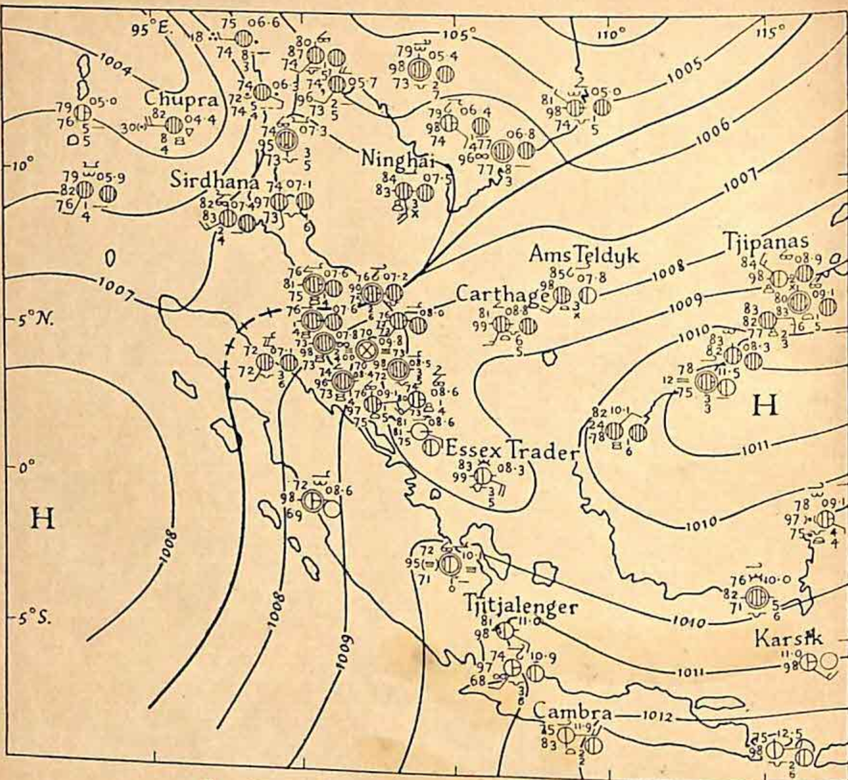


FIG. 88 Surface Chart, 0000 G.M.T., 29.7.49
(7.30 a.m. Malayan time)

of Medan. Thick altostratus is reported there and, no sign of great wind-shear being reported aloft, we may assume the altostratus to be related to a cumuliform development above the ground-level boundary. Therefore, in the absence of other evidence, we place the surface boundary also north of Medan.

There is little evidence for the position of the surface boundary over Malaya. Pressure falls regularly from Western Malaya northward into the Bay of Bengal, without a distinct pressure trough (Fig. 88). An aircraft crossing the northern Straits of Malacca during the night of 28th encountered considerable cumulonimbus about 5° to 6° N. We will accept this position as determining the surface boundary in the Straits.

The wind streams at 1000 ft. to 3000 ft. indicate the boundary to be somewhere near Kota Bharu. There is some justification for

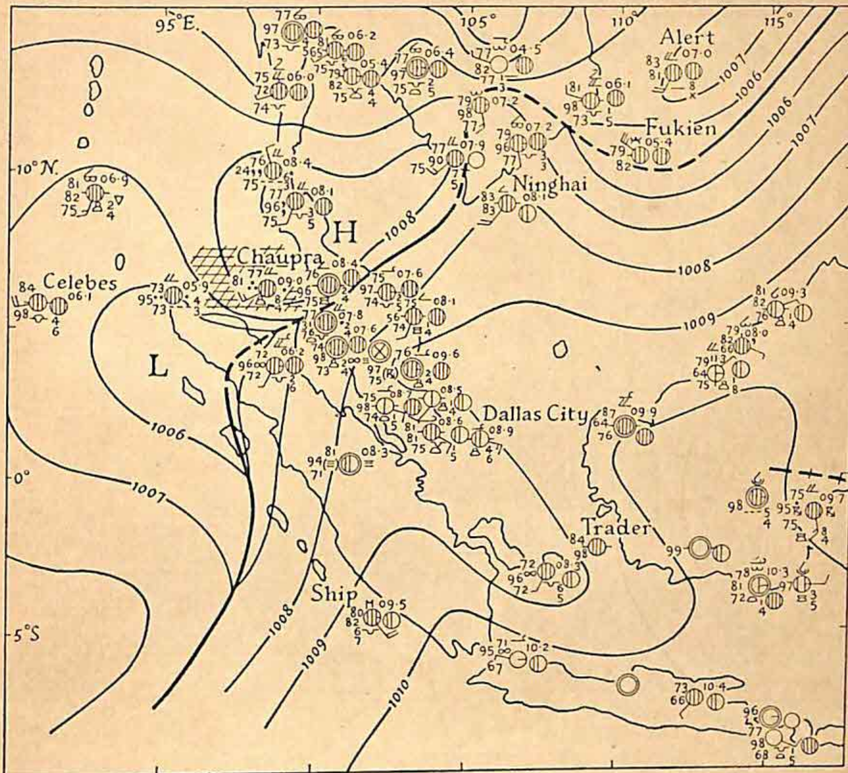


Fig. 89 Surface Chart, 0000 G.M.T., 30.7.49
(7.30 a.m. Malayan time)

placing the surface boundary to the north of the ship *Ninghai*, which reported a southerly wind; having no corroboration, we draw the surface boundary where it was well-established on the 28th, as shown in Fig. 88. No marked trough exists to fix its position over Eastern Malaya, though near Kota Bharu slight evidence of a pressure trough appears.

The wind analysis at 1000 ft. indicates that the minor air-stream boundary over Borneo is a little south of the Equator. Balikpapan reports a light easterly which, being contrary to the land breeze, suggests that the Trades blow across this station. Because Balikpapan reported rain in the distance, we may consider the surface boundary to be near by (Fig. 88).

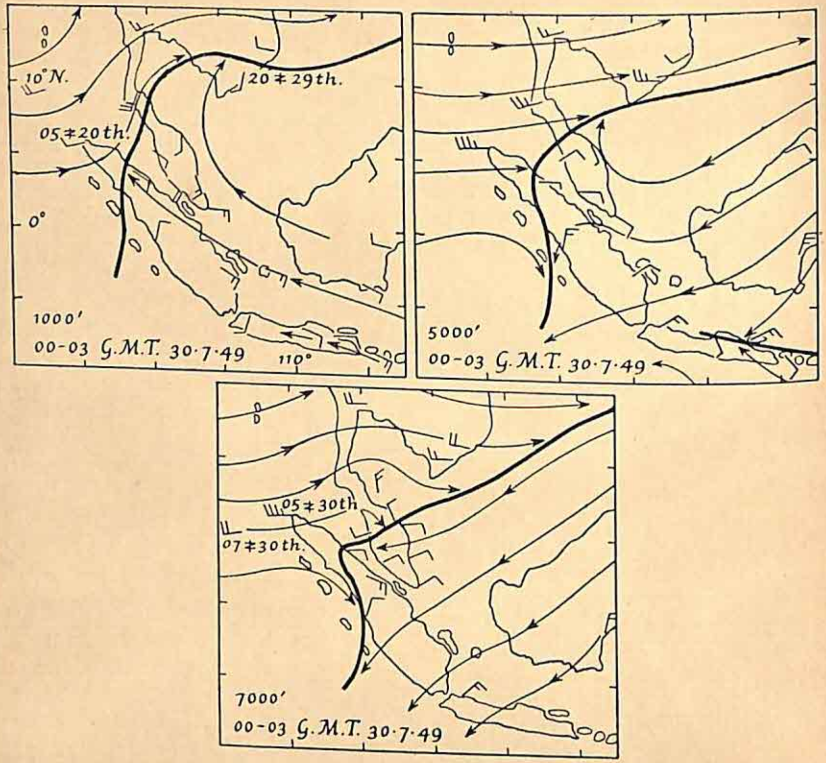


FIG. 90 Upper-wind Charts, 30.7.49

2. Situation at 0000 G.M.T. (0730 L.T.) on 30th July 1949

The analysis becomes more conclusive than for the previous day and confirms certain assumptions made then. A trough has developed through Penang and Kota Bharu (Fig. 89) in the position chosen for the Southern Boundary on the 29th and, although the Balikpapan thunderstorm might be a local disturbance, more probably it is due to the minor boundary.

Examining the changes aloft (Fig. 90), we find the greatest is at 7000 ft. where the Trades, oriented more northeasterly, have displaced the Southern Boundary well to the northwest (cf. 7000 ft. charts of Figs. 87 and 90). The minor air-stream boundary might be discernible over South Borneo were more observations available for the 30th.

There is no evidence for a substantial displacement in the 5000-ft. position of the Southern Boundary by the 30th, except that the

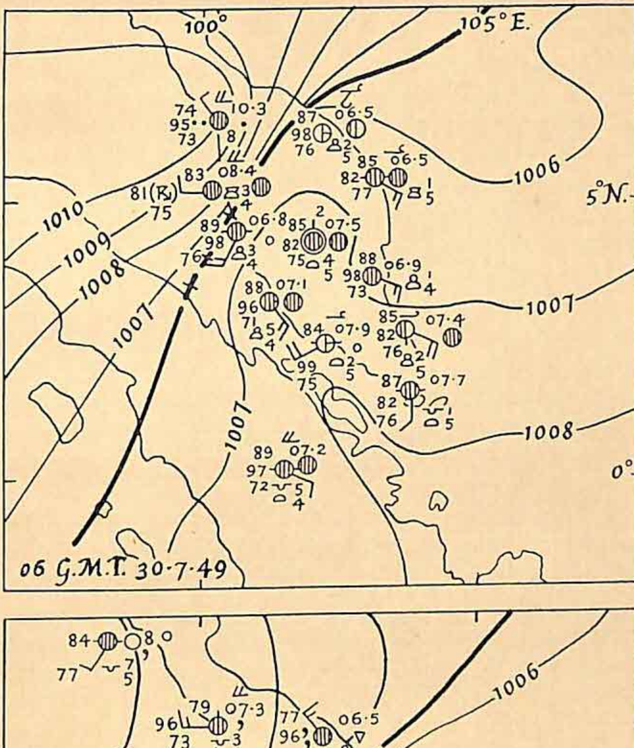


FIG. 91 Surface Charts, 30.7.49

northeasterly at Penang may be in the monsoon, in which case the boundary would lie south of Penang. The minor air-stream boundary lies over Java at 5000 ft., where the northeasterly and southeasterly Trades meet at a large angle.

At 1000 ft. (Fig. 90) the Southern Boundary is well defined by the contrasting directions of the Trades and monsoon (the 3000-ft. chart is similar). Southerlies reported on the Indochina coast are very significant because their direction could hardly be ascribed to local effects. Stream-lines for 1000 ft. cannot be drawn over Java and southeastern Borneo because directions reported there are so varied.

Projecting the Southern Boundary (Fig. 90) of 1000 ft. on to the surface chart (Fig. 89) is simple. Over Malaya a slight adjustment is needed and the boundary may be drawn in the pressure trough, supported by the secondary indication of thick altostratus and recent rain at Penang.

Farther west, the upper winds show that the boundary lies between Medan and Sabang without data to define it. Rain is reported at Sabang and in the northern straits by the 'Chaupra', where pressure is quite high. Drawing the boundary through the rain area would unwisely divorce it from the pressure trough, for which reason we place it near Medan on the evidence of altostratus cloud. The rain in the northern straits most likely resulted from convergence within the monsoon, evidence of which is found in the Sabang and Penang winds at 7000 ft. (Fig. 90).

3. Situation at 0600 G.M.T. (1330 L.T.) on 30th July 1949

During the morning of the 30th, the Southern Equatorial Air-stream Boundary has begun to move towards the southeast. The upper winds from 0600 to 0900 G.M.T. on the 30th (Fig. 92) indicate no new influences to cause it, although the actual advance is very evident at Kota Bharu, where winds at 1000 ft. and 3000 ft. changed from light southerlies at 0000 G.M.T. to strong westerlies by 1100 G.M.T. (cf. upper winds on Fig. 90 and Fig. 92).

A reason for the boundary displacement can be seen in the pressure changes which have occurred between 5° N. and 10° N. where pressure has risen so much from 29th to 30th that a weak low pressure has become a high (cf. Figs. 88 and 89). Probably the southward movement of the boundary has been brought about by the low-level outflow resulting from subsidence within this high, or by transfer of the high itself. By 0600 G.M.T. on the 30th, a

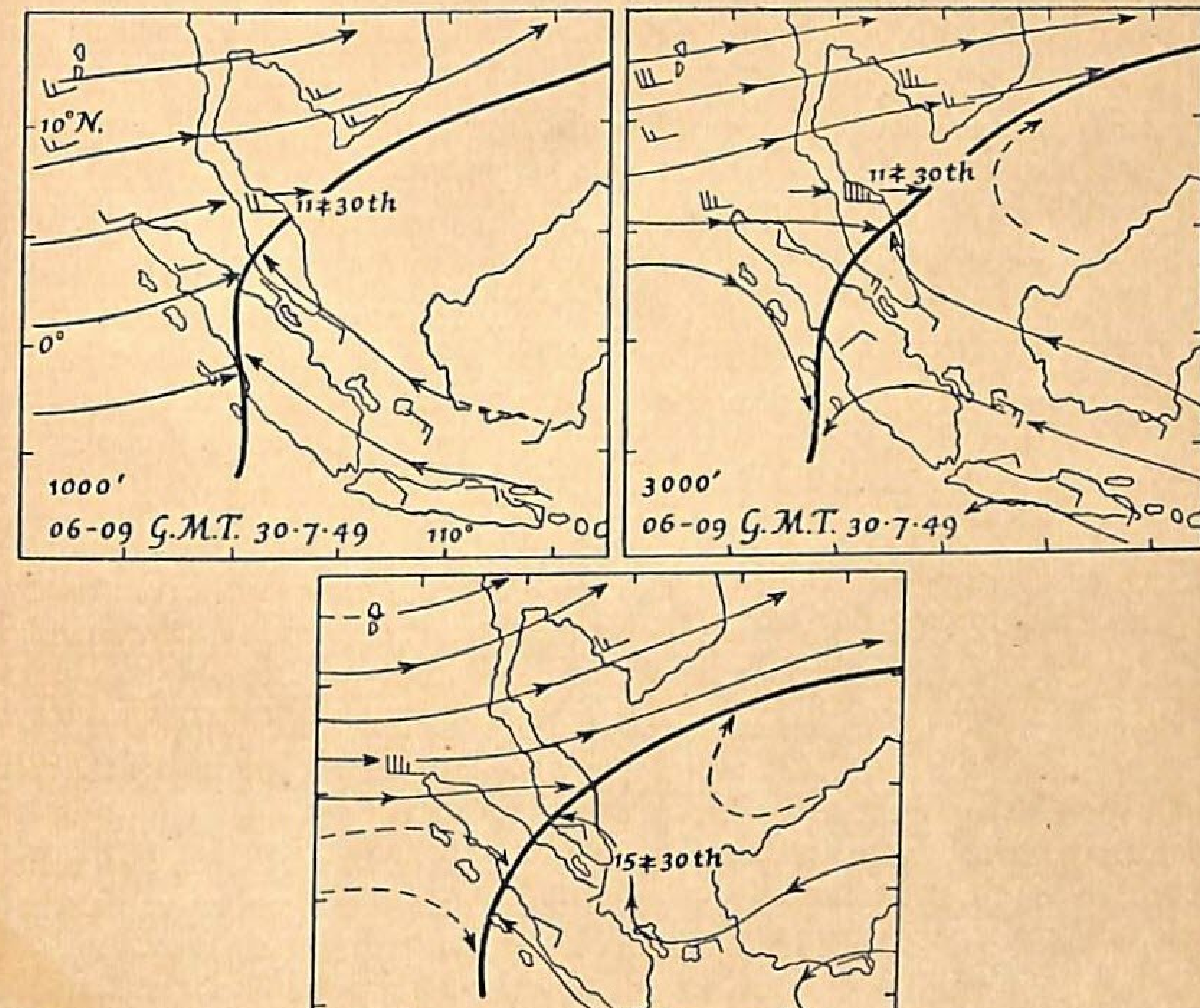


FIG. 92 Upper-wind Charts, 0600 to 0900 G.M.T., 30.7.49
(1.30 p.m. to 4.30 p.m. Malayan time)

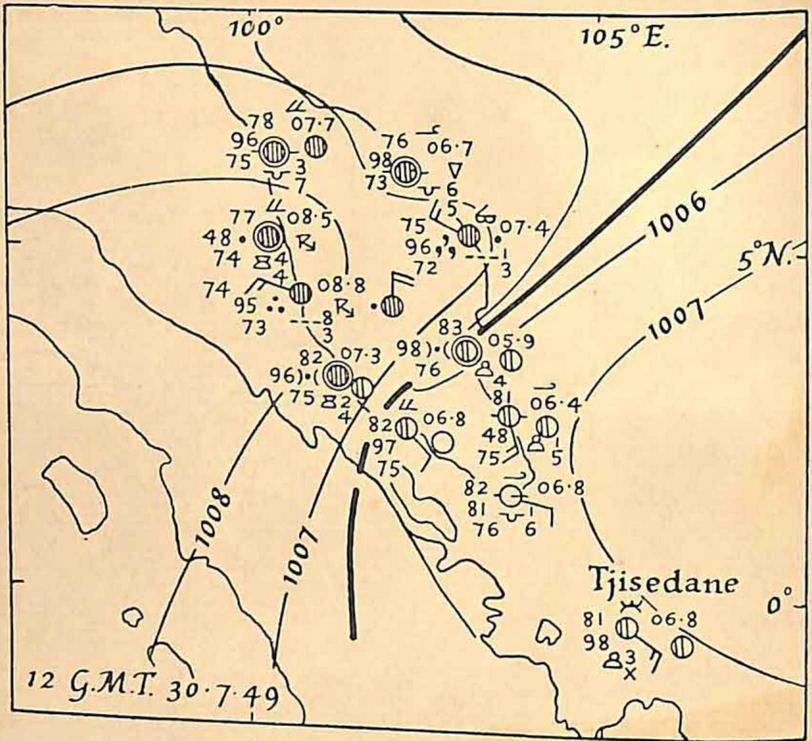


FIG. 93 Surface Chart, 30.7.49

thunderstorm developed at Penang (Fig. 91). Winds at 1000 ft. and 3000 ft. over Kota Bharu have turned to westerly (Fig. 92), and presumably these westerlies have come across Penang where rain prevented a balloon ascent. We must rely on the surface pressure trough to determine the position in the west, where the boundary may be deduced as lying near Sitiawan at 0600 G.M.T. (Fig. 91). Because the surface wind there is still from a southerly quarter, it is unlikely that the monsoon has travelled farther south.

4. Situation at 0900 G.M.T. (1630 L.T.) on 30th July 1949

By 0900 G.M.T. the position is clearer (Fig. 91). The boundary is crossing Sitiawan accompanied by a thunderstorm, and the monsoon flow is identified by the northwesterly surface winds at Penang and Kota Bharu. Once again the boundary is within the trough, while the axis of lowest pressure is now about 50 miles ahead of the

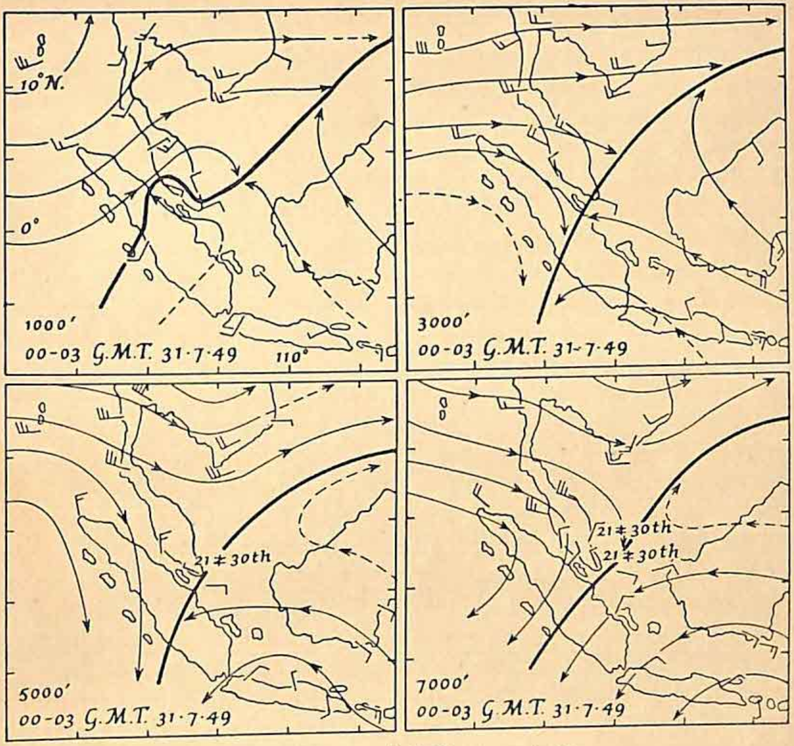


FIG. 94 Upper-wind Charts, 31.7.49

boundary. Rain is reported in the north. Over southern Malaya, where the surface winds are evidence of the Trades, there is no precipitation and little low cloud.

5. Situation at 1200 G.M.T. (1930 L.T.) on 30th July 1949

The progress of the boundary past Port Swettenham does not show in the surface winds at 1200 G.M.T. (Fig. 93). Because there are no further upper-wind reports, we must again rely on the pressure-trough method, remembering that the boundary tends to lag behind the axis of lowest pressure. We may then place the boundary close to Port Swettenham and Kuantan, both of which are reporting distant rain. Reports of southeasterlies and fine weather from Malacca and Kuantan and stations farther south denote that the Trades have not yet been displaced there. Rain is still widespread in the north, although clearing has begun at Alor Star.

6. Situation at 0000 to 0300 G.M.T. (0730–1030 L.T.) on 31st July 1949

The upper-wind charts for 5000 and 7000 ft. on the morning of the 31st show that the boundary has passed southeast of Singapore and that the monsoon stream, diverted to northwesterly or northerly, covers Malaya. The boundary does not lie far to the southeast because aircraft report that the Trades exist close to Singapore.

At 3000 ft. (Fig. 94) the Trades still lie over Singapore. At 1000 ft. the wind-field is more confused, because with the decreasing strength of the monsoon current, local Malayan winds are more prominent. Although the wind at 1000 ft. over Mersing shows the influence of the monsoon, the light northeasterly at Penang is probably a deep land breeze. Winds at Malacca and Singapore may be related to the Trade stream, a possibility which needs confirming on the surface chart of the 31st (not reproduced).

At 0000 G.M.T. on the 31st a sheet of altostratus lies across central Malaya. Lacking reports of rain or cumuliform cloud even on the west coast, it appears unlikely that the altostratus sheet is related to the boundary.

Slight thunderstorm activity is reported at Mersing during the afternoon. No upper-wind reports being available, we cannot determine whether the thunderstorm is associated with the boundary. Winds at Singapore vary between northerly and southeasterly during the 31st with periods of thunder and rain. The boundary, marked by a broken line of cumulonimbus, may lie across that region and the thunderstorm at Mersing may mark its northernmost edge of activity.

7. General Considerations

The examples we have described illustrate the difficulties encountered in analysing low-latitude charts. The difficulties may nearly all be traced to the scarcity of reports and to unrepresentative individual observations. To determine the air-stream boundaries, the analyst must weigh the relative importance of wind contrasts at low levels, of the pressure trough and of the distribution of rain. Since much rain occurs wholly within a stream and remote from the boundary, and since many boundaries are either outside the major troughs or at a distance from the axes of low pressure, the analyst must always prefer the position indicated by the wind streams.

On many days, however, the winds may be so light that the streams are difficult to trace. Then there is little alternative except relying on a position for the boundary suggested by the pressure

trough or the distribution of rain. Frequently the trough is not distinctly seen and the precipitation not localised, so that the boundary might be in one of several positions. The selection of one will be influenced by the need for continuity from one day's chart to the next, and by avoiding boundary movements which are excessively fast in relation to the observed wind components.

The analysis of the equatorial weather map cannot be conducted with the same precision as the analysis of a chart of temperate latitudes. Increased reports of wind and weather may facilitate it, though the absence of useful pressure-wind relationships at low latitudes will still constitute a serious drawback. The present system must be considered as only an expedient ultimately to be replaced by a method based on new and more rigorous theories.

CHAPTER XIII

Local Disturbances

The previous chapter (XII) examined in detail the synoptic situation from 29th to 31st July 1949, an example of a condition relatively simple to analyse. During much of the year, weather analysis is complicated by pressure troughs too weak to be determined from the observations of Southeast Asia's scattered stations. Winds are then frequently much lighter than those of the example in Chapter XII, and although the boundaries at upper levels may be located by examining the wind-fields, the variations of wind direction are so small that they make analysis at the ground very difficult.

At low levels, land and sea breezes may mask the general flow near the major land masses. It is therefore necessary in assessing the significance of a low-level observation of wind, to know the normal land-breeze or sea-breeze direction which would occur in a locality if the air flow were negligible. Only after subtracting from the actual wind observation any component which could be ascribed to land-breeze or sea-breeze effects, can the analyst assess the overall flow. To estimate the significance of each observation of wind involves plotting, for each station, the wind observations at each hour or half-hour throughout the day; then the effect of local breezes becomes evident (Fig. 51).

1. The Sea Breeze

When no well-defined stream flow of air exists, the sea breeze is brought about by the difference between temperatures of land and sea during insolation. Only a small part of solar heat contributes to raise the temperature of the sea; most of it evaporates water from the surface, and some is reflected. The surface temperature is raised a little, for which reason shallow inland waters are sometimes warmer by day than by night, but any slight rise of surface temperature at sea is soon dissipated downwards by wave motion.

On land, some solar heat may be reflected and a little may evaporate ground water. Most of it goes to raise the temperature of the ground, which is a poor conductor, so that heating is restricted to its surface. The air in contact with it is heated and, each day

from sunrise onwards, a gradient of temperature is built up across a coast-line, with the higher temperatures over the land. As this temperature gradient is built up, if no immediate exchange of air takes place by convection, there will be no immediate change of pressure at the ground (p_o of Fig. 95), where pressure is equivalent to the weight of the overlying atmosphere and could vary only if air were transported. The air in the lowest layer is warmed and expands upwards, so that the difference ($p_o - p_B$) between pressure near the ground and pressure at height B (Fig. 95) decreases. Pressure (p_o) near the ground remains constant, so that (p_B) at the upper level increases.

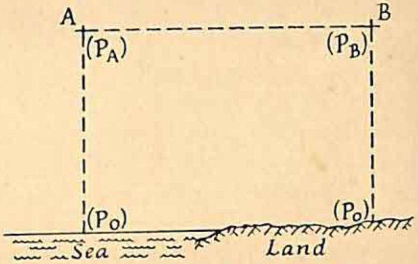


FIG. 95 Formation of the Sea Breeze

If the distribution of pressure with height over the sea was initially the same as that over the land ($p_A = p_B$), the first effect of solar heating would be to create a slight gradient of pressure at right angles to the coast at an upper level, so that pressures would become greater over the land than over the sea ($p_B > p_A$). Such a difference would cause a gentle flow from land to sea, extending through a deep layer, although not very noticeable in the air nearest the ground. The outflow lowers the ground pressure below that over the sea, and eventually causes a compensating low-level flow from sea to land—the sea breeze. It normally sets in first near the coast-line, though sometimes at a distance from the coast where the upper seaward-flowing current subsides to return landward at lower levels.

The sea breeze frequently begins at different times along the coast, usually as a gentle wind of 5 m.p.h. blowing at right angles to the coast and, if the general flow aloft exerts no influence, increasing to 15 or 20 m.p.h. by the afternoon and decreasing again at sunset. The afternoon increase of speed involves a deepening of the sea breeze to about 1000 ft. above the ground.

The sea breeze may sometimes begin nearer midday, particularly when the main flow of air aloft is from land to sea. Delayed sea breeze has special interest in tropical countries because it may take the form of a line-squall oriented parallel to the coast. The initial upper seaward flow subsides at a considerable distance from the coast and the low-level return current is retarded by the general flow of the air stream, so that by the time the sea breeze arrives its

temperature is lower than that of the air over the land which has warmed by contact. Hence the sea breeze arrives with a structure resembling a cold front of middle latitudes; cold air is undercutting warmer, favouring the development of convective cloud and a line-squall.

2. The Land Breeze

At night, both land and sea radiate heat: much is absorbed by the atmosphere, some reflected back by clouds and an appreciable amount lost into space. The ground temperature falls rapidly after sunset, particularly if there is no cloud, and this continues until sunrise. Because heat is conducted very slowly through the earth, cooling is confined to its upper shell. The sea surface also radiates heat, but its fall in temperature is negligible.

The result in the night is that cooled air at low levels flows from the land towards the sea, forming the land breeze which, provided the sky is clear, starts an hour or two after sunset and continues until daybreak. Because it depends on cooling of the ground, the land breeze is confined to a depth of not more than a few hundred feet.

The nocturnal seaward flow is not restricted to the coast. Inland air on the hills is also cooled by radiation and gently drifts down the hillsides and valleys. Where there are no easy outlets the air gathers in pools, but it mostly emerges on the coastal plains to reinforce the land breeze. This flow is termed a "katabatic wind" which, although normally light, may become strong if funnelled in a narrow valley.

On occasions the land breeze, like the sea breeze, may begin as a squall, especially when its development is retarded during a general air flow from sea to land. The prolonged cooling over the land causes low temperatures there, although air over the sea maintains much the same temperature as in daytime. When the contrast of temperature (and consequently of pressure) is sufficiently great, retardation by the general flow is overcome. The cooled air then flows seaward to undercut the warmer air and, if the latter is unstable, a line squall may result.

Whether or not the land breeze will develop is critically governed by the general flow: if the latter is very strong, a land breeze will not usually develop because the lower air is swept away with the main air current before becoming greatly cooled by the ground.

Line-squalls and convective phenomena formed offshore when the land breezes undercut the sea air are often vigorous. Neumann[12] has shown that cumuliform development and nocturnal thunderstorms on the coast depend on the convergent or divergent nature of

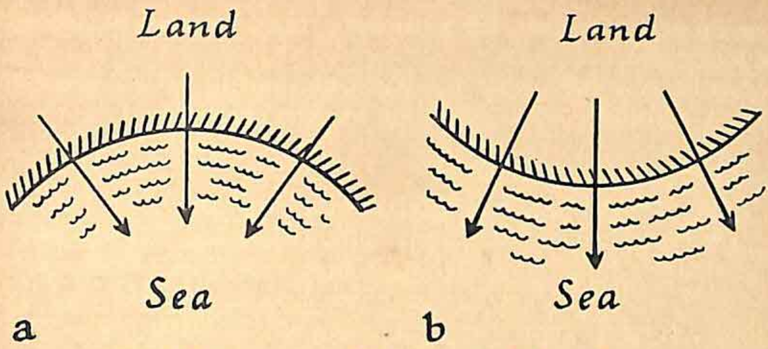


FIG. 96 Convergence and Divergence in the Land Breeze

the land breeze, which in turn depends on the curvature of the coast-line.

When the general flow of air is negligible, nocturnal convection offshore is easily established. If the coast is concave towards the sea, the land breezes, being at right angles to the coast, converge towards the centre of the concavity (Fig. 96*a*), concentrating the seaward flow to undercut the sea air and create a line-squall and clouds. If, on the other hand, the coast is convex towards the sea (Fig. 96*b*), the land breeze is divergent and, although undercutting and convection are still possible, they will be more dispersed than on a concave coast.

3. Line-squalls

Although winds are frequently light before and after the passage of a slow-moving boundary, they may at some point become a vigorous squall. If a boundary squall occurs at a reporting hour, the location of the boundary is readily determined, but should it occur between reporting hours the position of the boundary will not be revealed. Reports at each observing station must be watched for signs of squall passages, and stations should have a system for announcing squall warnings to a co-ordinating centre even outside the usual reporting hours.

This does not mean that, in the absence of a land breeze or sea breeze, every squall passing a station denotes the passage of an air-stream boundary; isolated, stationary convective clouds are often accompanied by sharp squalls. When many simultaneous squalls from the same direction are reported at various stations in a line across a good observational network, the probability of their being due to isolated cumuli is low. If the upper winds show that a bound-

ary lies across an area, the squalls may often be taken to indicate its movement.

Let us now investigate the particular squalls of the Malayan region, classifying them according to their origin.

4. Line-squalls of Malaya*

Because large changes of wind direction are common in the slack air streams of low latitudes, a squall needs defining. To merit attention, the winds initiating the squall must be of greater than normal strength. The autographic records show that the average wind over Malaya is rarely more than about 18 m.p.h. and that gusts are not consistently over 30 m.p.h. except at hill-stations. Where gusts exceed 30 m.p.h., they are usually associated with marked changes of wind direction and large abrupt increases of wind speed. In this investigation, 'significant squalls' will be those in which the wind after a directional change contains gusts of over 30 m.p.h.

In places, most gusts of this speed are accompanied by significant squalls. For instance, at Malacca on 54 occasions per year gusts exceed 30 m.p.h. and on 49 of these a marked squall accompanies them. At hill-stations the ratio is not so high because the general flow aloft is stronger than at lower levels.

Isolated cumulonimbus often occur with vigorous, local, non-linear squalls. It is desirable to determine how many of the squalls at an observing station are line-squalls and how many are caused by isolated cumulonimbus. This involves examining the squalls passing each Malayan station on certain days and, after plotting the times of squall passage (recorded by autographic anemometers), to draw 'isochrones' or lines of equal time of squall passage among the stations. If the isochrones of reported squalls appear as straight lines or smooth curves progressing across the country at a speed approximating to the squall wind, it is reasonable to suppose that they represent the varying positions of one moving line-squall.

Line-squalls will not necessarily be composed of continuous lines of cumulonimbus with equally strong winds at every point along them, because the roughness of the terrain influences both cloud development and winds. Observing stations sheltered from certain directions may not record a significant squall, even though the line-squall is indicated by neighbouring stations. To present a complete picture, some minor wind changes must be plotted. When a line-squall is established from synchronous observations of significant squalls, the time of any lesser wind change at another station must

* Much of this section has been published in *The Malayan Journal of Tropical Geography*.[67]

be plotted, provided the lesser change is apparently not due to a local effect. Thus a weak wind change inland during the afternoon when convection is at its greatest should not be considered, but if two coastal stations record significant squalls from the sea about midnight, an adjacent weaker wind change of similar sense might be symptomatic, being contrary to the normal land-breeze régime.

Accompanying the movement of the boundary of the 30th July 1949 (Chapter XII), many vigorous squalls were reported, so that it seems likely that the boundary passed as a true line-squall. The squalls of 30th July are plotted in Fig. 98 in the following manner. One wind-arrow at a station shows the direction of the wind *before* the squall, with the speed in m.p.h. noted at the end of it. The other arrow shows the wind direction immediately *after* the change, the average Beaufort Force within the squall being shown by feathers on the arrow, while the value of the peak gust in m.p.h. is plotted to the left of the station. The local time of the wind change is given as four figures above the station, and below it are plotted the temperature drop and the rainfall, each within a certain number of minutes of the time of the squall passage. Thus at Port Swettenham on 30th July 1949 (Fig. 98), the wind changed at 1930 hours L.T. from southerly 3 m.p.h. to north-northwesterly Force 7 with gusts to 39 m.p.h. Temperature fell 8° F. in 30 minutes and there was a fall of 0·06 inches of rain in 10 minutes.

The isochrones appear to show that the passage on this day was in the form of a true line-squall. The first report of it was from Alor Star at 1025, which was not very much later than the normal start of sea breezes, but its gusts of 33 m.p.h. were abnormal for sea breezes and the rain was exceptionally heavy. Taking also into account that the sea breeze is usually from a more westerly direction, the Alor Star wind change appeared to indicate a line-squall, possibly influenced or introduced by the sea breeze.

Two squalls occurred at Penang, the one at 1305 and the other at 1400 hours. The sea breeze had already developed before 1305, so that, although winds were stronger after the second change, the first change was probably the more important evidence of a line-squall. On the other hand, the line-squall, if it existed, was possibly an uneven broken line, making both changes relevant. This likelihood was supported by a report from H.M.S. *Mendip* (anchored at Penang) that the storm broke at 1235 when wind rose to Force 5 within a few minutes.

There was little local indication at Penang of the approaching storm other than thickening cloud to the west. The storm caused considerable damage, many lighters being swept away from mer-

chant ships. The Singapore *Sunday Times* of 31st July gave the following account of the storm: 'A gale of almost cyclonic force which hit Penang was followed by heavy seas. . . . Ground-floor rooms were flooded, and the seas that crashed against the sea-wall catapulted in places to 60 ft. in the air. Streets facing the waterfront were flooded most of the afternoon. The early wind crashed trees across streets all over the island of Penang. Another storm, apparently connected with that at Penang, hit Ipoh later and lasted for about an hour.'

The upper winds over the period 29th to 31st July 1949 are shown with local times in Fig. 97. At 0730 on 29th, southerlies (obviously Southern Hemisphere Trades) covered Penang to a depth of 5000 ft., while at higher levels were the westerlies of the Indian Southwest Monsoon. By 31st, northwesterlies had descended to ground-level because the monsoon had been diverted around the northern tip of Sumatra at low levels and was driving southwards the Southern Hemisphere Trades and the Southern Equatorial Boundary.

Surface-wind changes at Sitiawan and both surface and upper-wind changes at Port Swettenham support the conception of a line-squall moving southwards at about 20 m.p.h., having regard to the reported wind-speeds of 20 to 30 m.p.h. At Kota Bharu there was an abrupt fall of temperature, too large to be ascribed wholly to the light shower, but unfortunately no anemometer was functioning. Changes of upper-wind direction indicated a change of air stream (Fig. 97), and the wind of 45 knots at 3000 ft. suggested the passage of a severe squall.

Inland changes were more difficult to interpret. The squall struck Ipoh from the southwest; the town is well shadowed by hills to the northwest but open to the southwest. Kuala Lipis, which experienced a marked squall, is entirely surrounded by hills, and it is improbable that any reported direction could be representative of the squall.

Farther south, at Temerloh, Malacca and Mersing, changes were slight and not necessarily connected because the direction of the changes were in accordance with the normal night sequences. Vigorous squalls could not be expected in the south, because a comparison of the morning upper winds of 31st at Penang and Port Swettenham (Fig. 97) showed that the force of the squall was probably being diminished as it progressed southwards.

Details in this example show that the changes of wind and weather at stations may offer significant information of the progress of fast-moving boundaries, but many travel too slowly to be traced by this

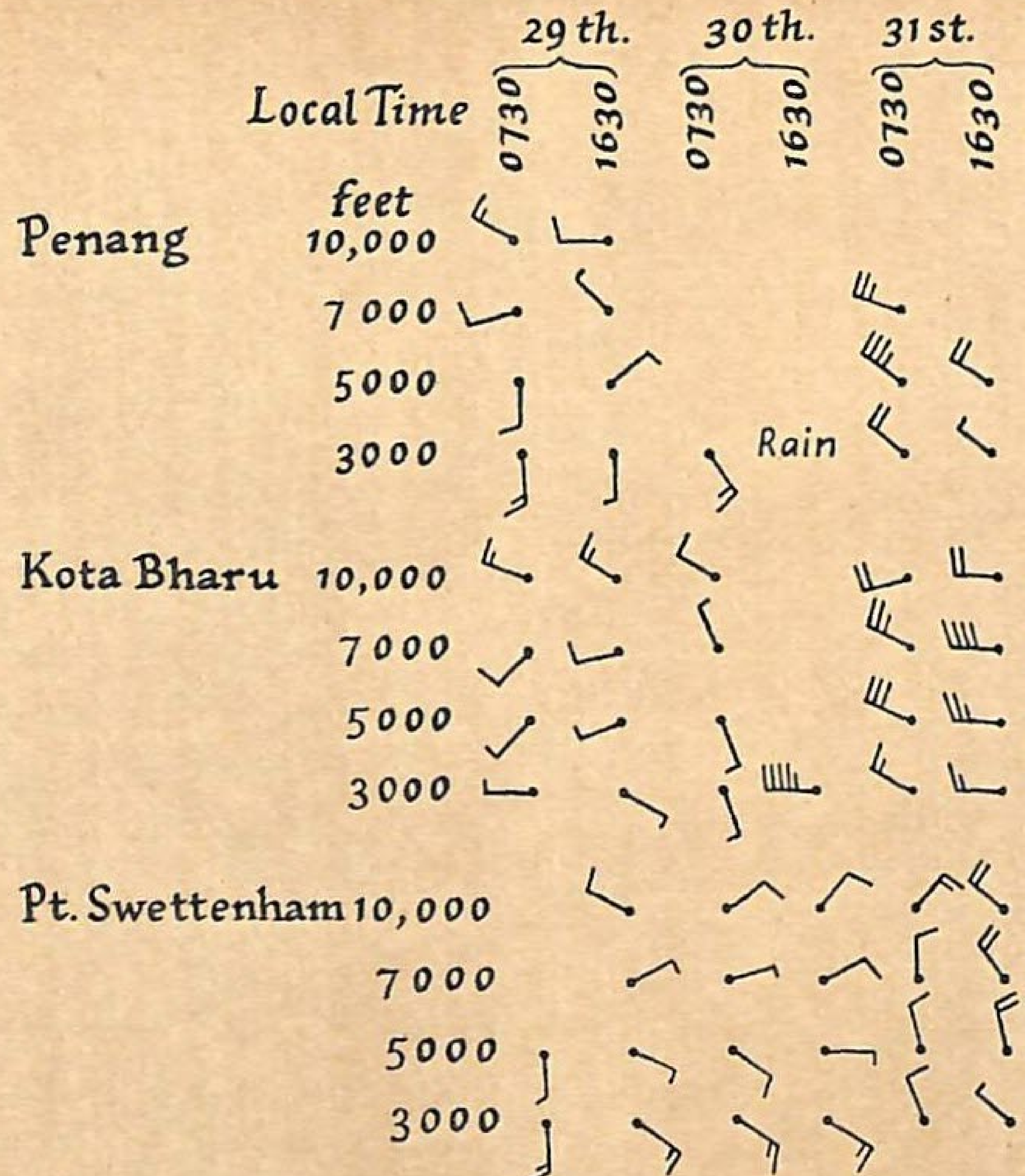


FIG. 97 Upper Winds, 29.7.49–31.7.49

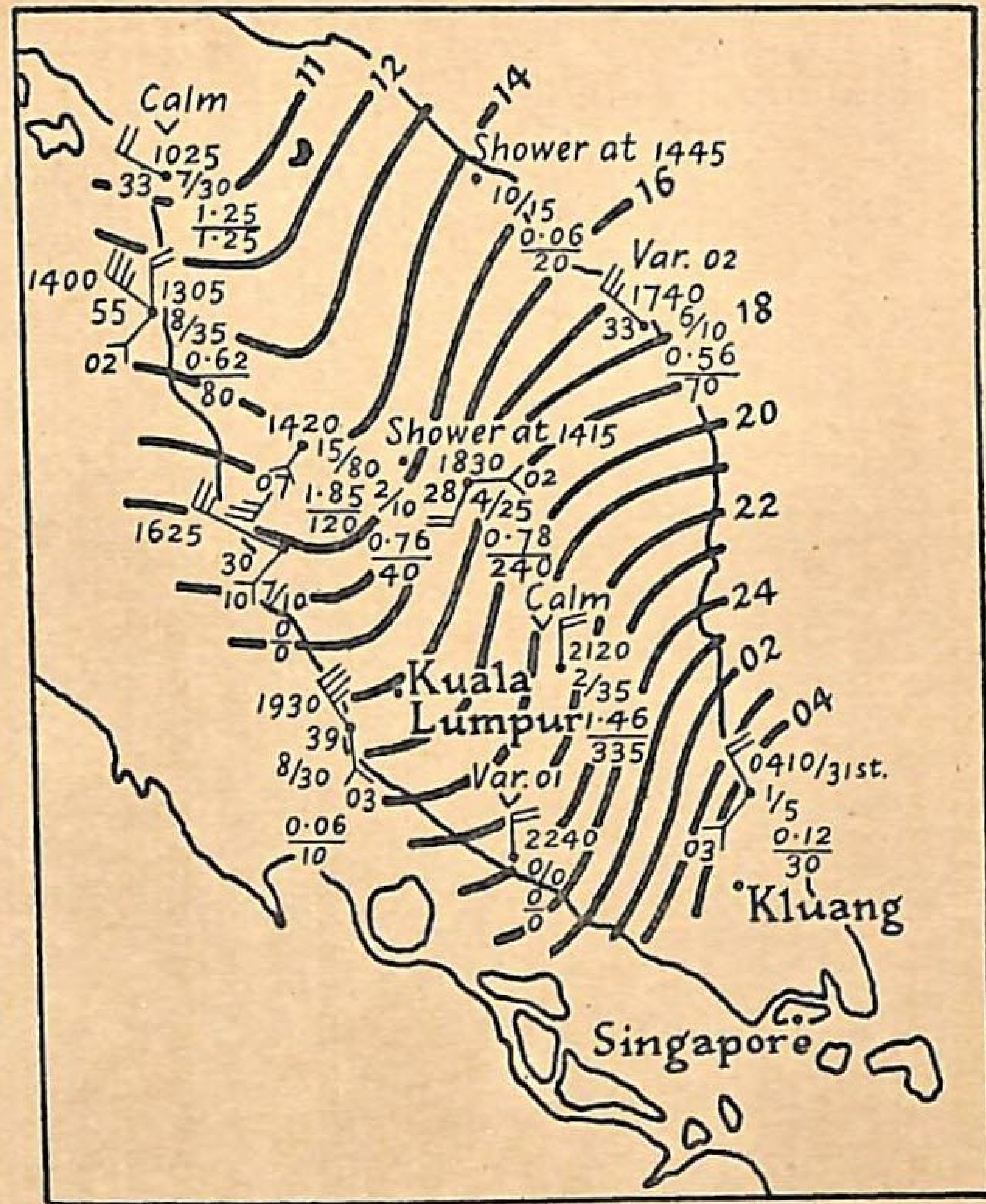


FIG. 98 Squall of 30.7.49

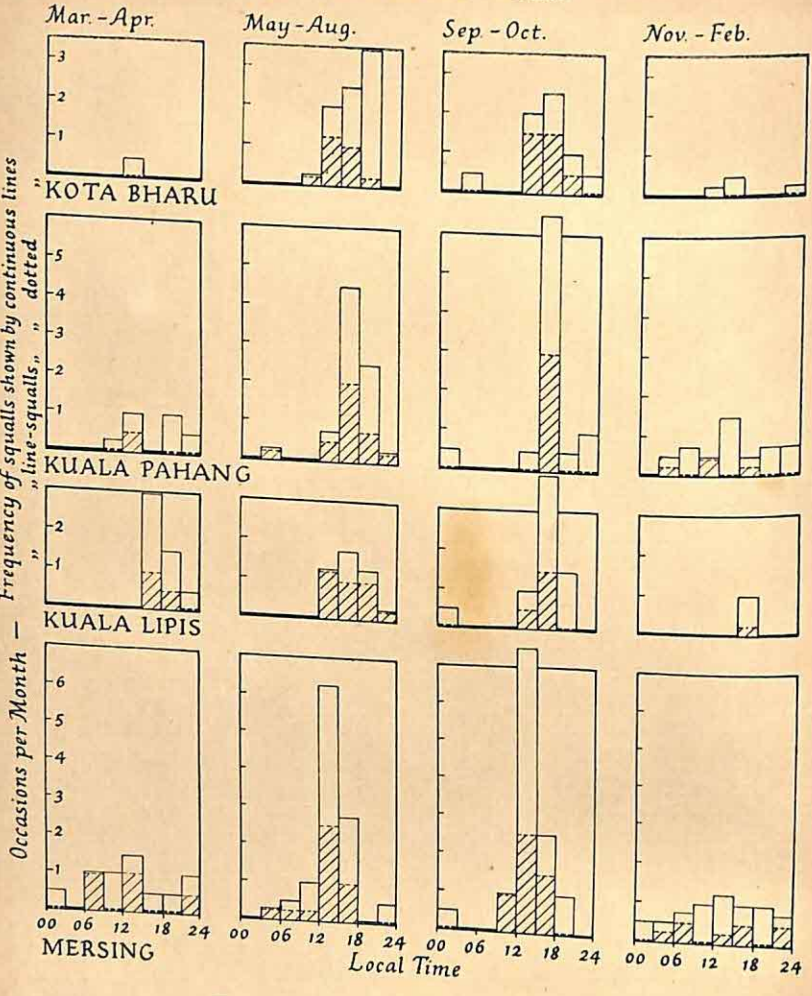


FIG. 99 Frequency of Squalls—Malaya

method. In a region where winds are generally light, the occurrence of an occasional strong wind is readily noticed and the tracing of a line-squall is simplified. At low latitudes, the synoptic method of analysis should be supplemented by a system in which 'significant squalls' are reported and plotted to determine linear characteristics, whether or not related to true boundaries.

Stationary local thunderstorms frequently produce squalls, and a squall-reporting network must be close before moving line-squalls

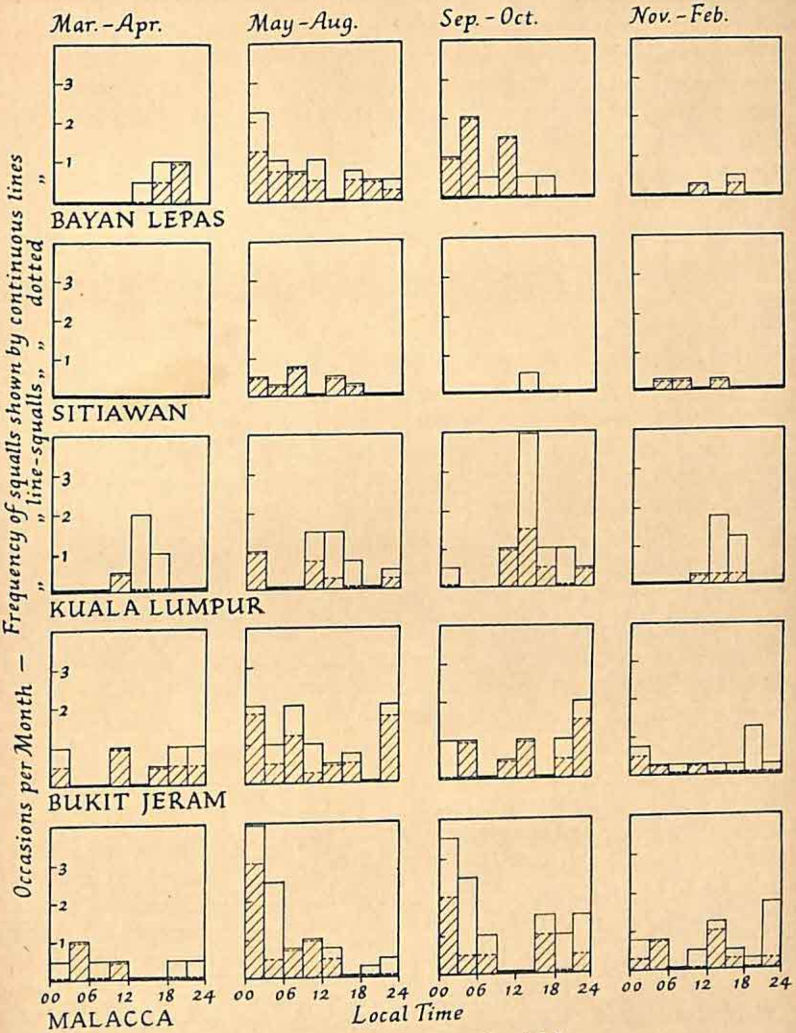


FIG. 100 Frequency of Squalls—Malaya

can be distinguished from isolated convectional storms.

The relative frequency of line-squalls and 'significant squalls' at Malayan stations will now be examined for the years 1936 and 1937, when there was a well-distributed network of sixteen Dines recording anemometers in the Peninsula. Every 'significant squall' (with gusts of 30 m.p.h.) passing each station was noted. For each day a chart has been plotted (in the manner of Fig. 98), showing the times

when squalls passed. Wherever possible, isochrones have been drawn. This method at first appears open to the criticism that, on a single day under some conditions of atmospheric instability, one station might experience several local thunderstorms vigorous enough to indicate 'significant squalls' and render the isochrones unreliable. Such was not the case: two squalls rarely passed a station in one day, so that the question of arbitrarily choosing one did not arise. When any squall passed at a time of day unfavourable for convection, it could be accepted for drawing the isochrones of squalls.

On Figs. 99 and 100 are the average numbers of 'significant squalls' and the average numbers of possible line-squalls each month passing each station during fixed hours of the day. The frequency of 'significant squalls' is denoted by the height of the open columns, and that of probable line-squalls by the height of the hatched columns bounded by dotted lines. Thus at Kuala Lipis (Fig. 99) between 1500 and 1800 during March and April, there are on the average three significant squalls and one probable line-squall each month.

Squalls in the Northeast Monsoon

These diagrams reveal that squalls are rare during the Northeast Monsoon (November to February), and that on the east coast and inland (Fig. 99) only a very small proportion could be classed as line-squalls. The highest frequency of squalls at this season is in the early afternoon as local convective storms.

A few weak line-squalls may occur during this monsoon, as in the case of Fig. 102 where isochrones may be drawn. They are not easy to justify because the speed of travel in the central section would be about 35 m.p.h., which exceeds that of the observed winds. Wind changes near the west coast on this day may have marked the beginning of the land breeze, although it is not easy to understand how the squall at Frasers Hill could have occurred under local influences alone. Low-level wind observations at Kuala Lumpur and Singapore (Fig. 101), however, show a general change from light northwesterlies to moderate or strong northerlies and north-easterlies when the squall passes, giving reason to believe that whether truly linear or not, it was related to a moving convergence zone within the monsoon. At the time of passing, air which had previously been flowing down the west side of the Malayan mountains was diverted to a track more consistent with the monsoon.

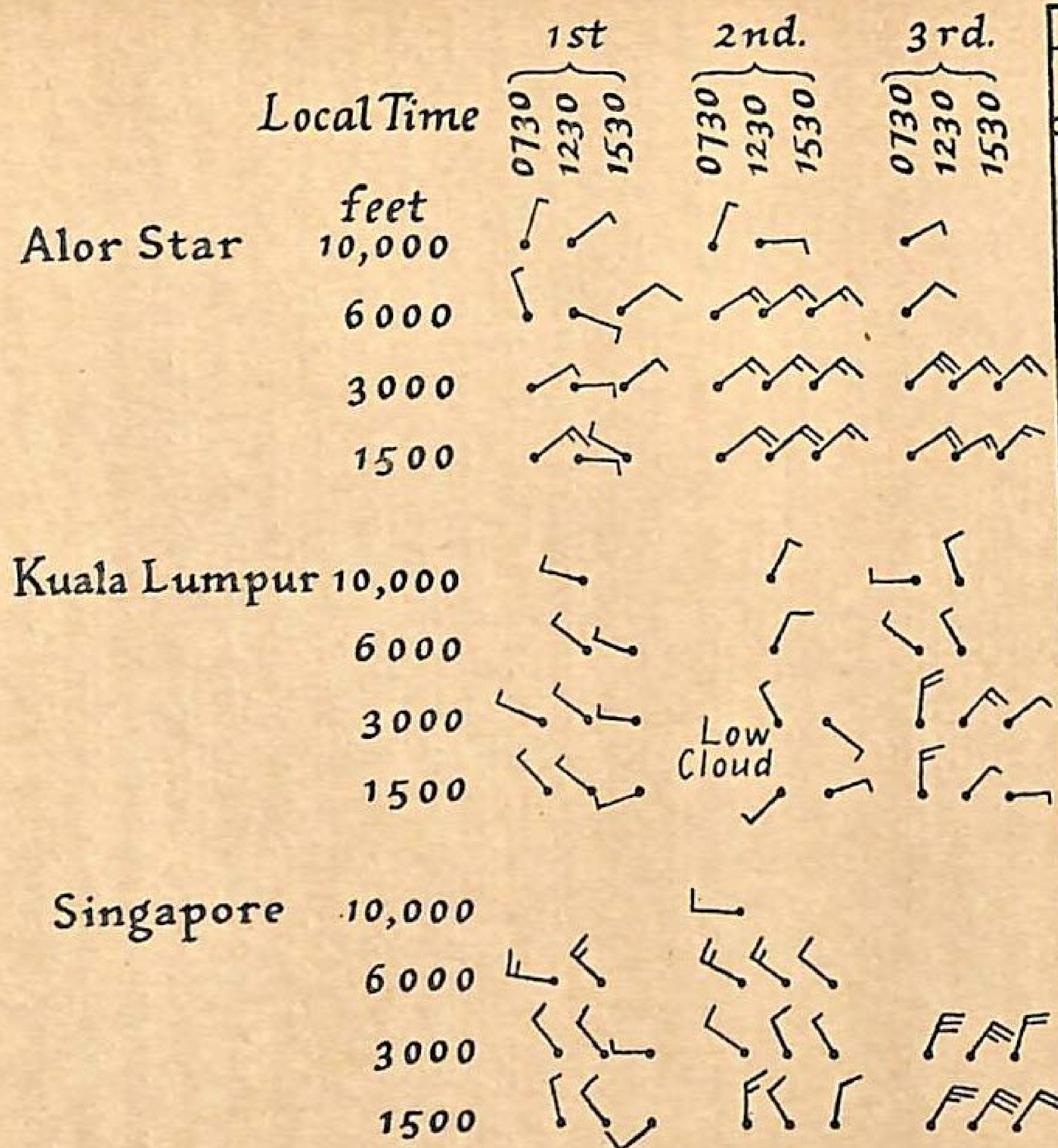


Fig. 101 Upper Winds, 1–3.1.37

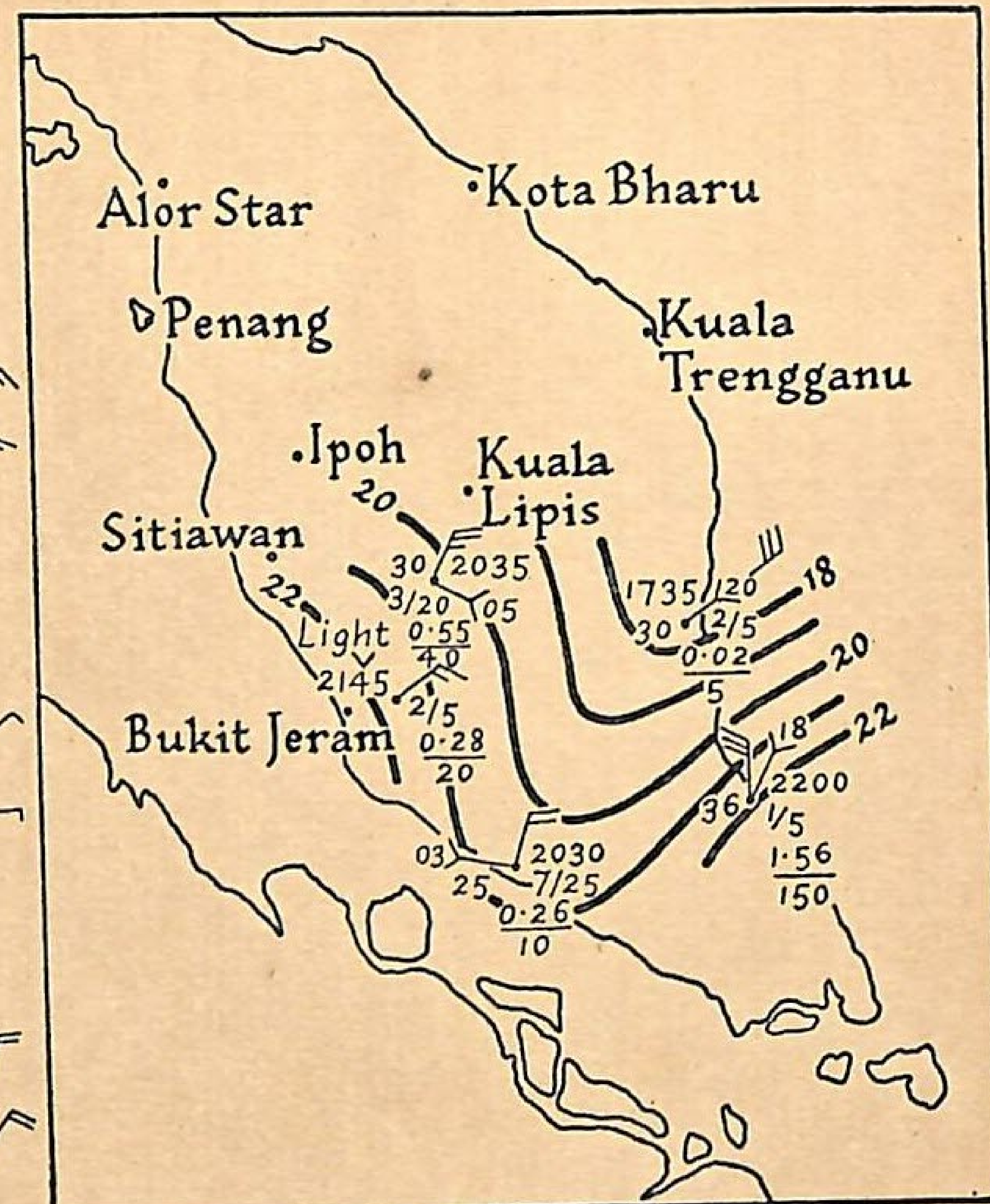


Fig. 102 Squall of 2.1.37

Line-squalls from the Southwest with Change of Air Stream

Another point brought out by the frequency diagrams of Fig. 100 is that many early-morning squalls at Kuala Lumpur, Bukit Jeram and Malacca at the time of the Southwest Monsoon (May to August) are line-squalls, as in Fig. 104. On the 8th June 1937, a well-defined line-squall crossed Western Malaya and moved northeastward at about 20 m.p.h. The light low-level winds at Kuala Lumpur and Singapore were southerly to southwesterly before it passed (Fig. 103), and because it is unlikely that air with this orientation could have crossed the 8000-ft. Sumatran mountains, it appears reasonable to argue that the flow should consist of Southern Hemisphere Trades.

Winds over Singapore changed to strong westerlies after the squall (Fig. 103), which could have resulted from a change of direction wholly within the Trades, but more likely from Southwest Monsoon air which had crossed Sumatra. Changes in the upper winds over Kuala Lumpur supported the analysis that the southerlies of the Trades had been displaced by a new stream from the northwest, possibly a low-level diversion of the monsoon air around the northern tip of Sumatra similar to the change depicted in Fig. 98. Although the squall introduced little rain or temperature fall to the north, there was a heavy shower at Singapore.

Line-squalls from the Southwest with Little Change Aloft

Most line-squalls from the southwest during the Southwest Monsoon have a very different character and negligible changes in wind structure aloft, as in the example of 1st July 1936 (Figs. 105 and 106).

The winds above Kuala Lumpur and Singapore up to 3000 ft., both before and after the passage, were undiverted Trades from the south. While the west-southwesterlies at 6000 ft. may also have been Trades, more probably they were monsoon air which had crossed Sumatra, because directions of flow at 10,000 ft. definitely show that origin. Some convergence might be suspected in the reported veer of wind and slightly increased velocities at 3000 to 6000 ft. over Kuala Lumpur, but it is unlikely that this change contributed to the formation of the line-squall: many similar squalls showed no such changes aloft.

Line-squalls of this type, lying wholly within a single air stream, commonly strike the coast between Port Swettenham and Singapore and they are known locally as 'Sumatras'. They consist of a practically continuous line of towering cumulus or cumulonimbus,

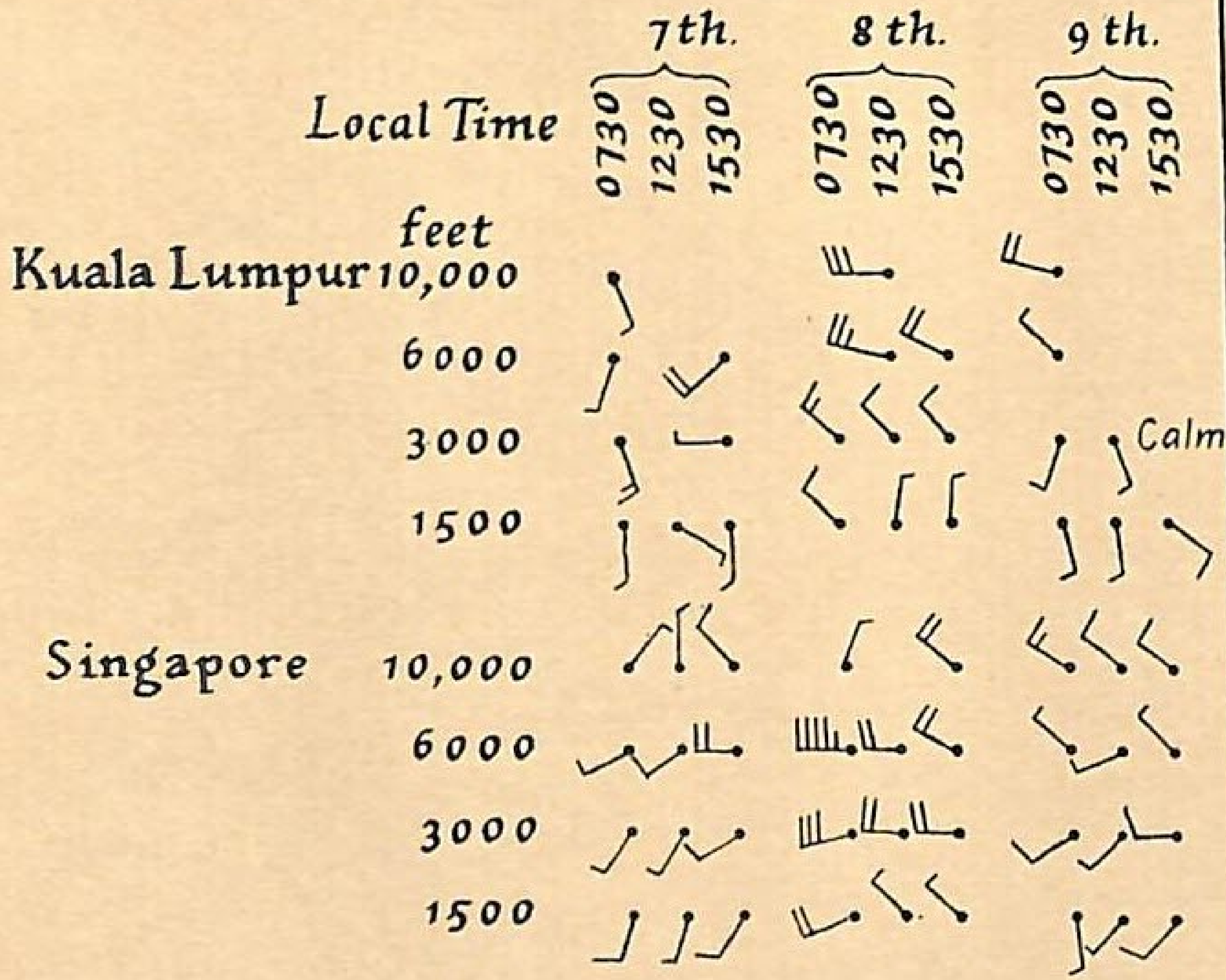


FIG. 103 Upper Winds, 7–9.6.37

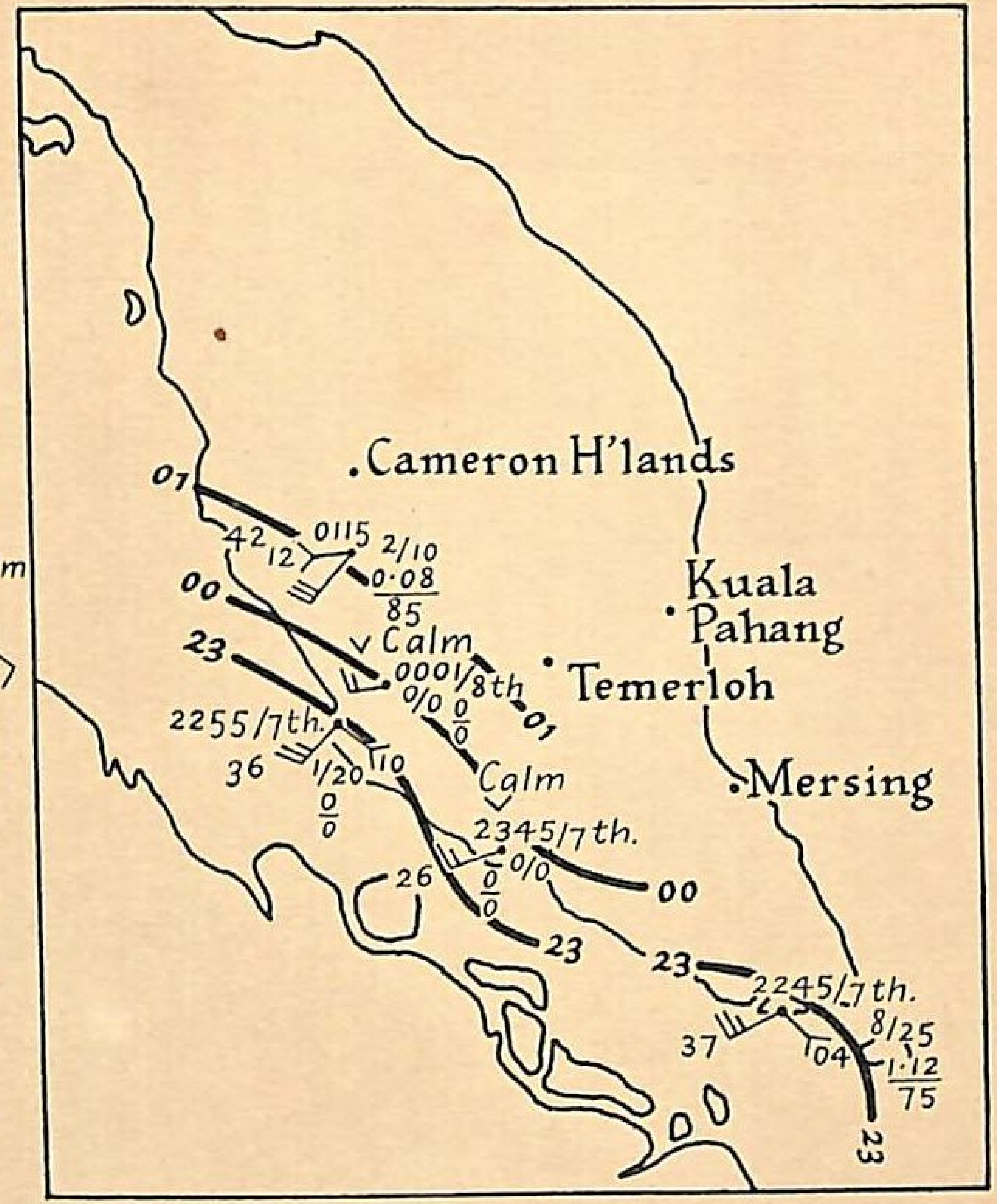


FIG. 104 Squall of 8.6.37

except occasionally when they may be composed of a few isolated storms along an uncertain line. 'Sumatras' are mainly confined to the period May–September, and some similar line-squalls between March and November may not be so classified because they accompany a moving air-stream boundary.

On striking the west coast, a well-developed 'Sumatra' is accompanied by gusts of over 50 m.p.h., and by squalls decreasing in violence as the line moves inland at about 30 m.p.h. Most occasions when squalls cross coastal stations involve temperature drops of 5° F. within five minutes or so, a decrease not wholly due to cooling by rain because it occurs even without precipitation, a condition not found when a line-squall accompanies boundary movement.

'Sumatras' always appear at night or in the early morning. During 1936–7 the greatest number crossed the west Malayan coast either between 2300 and midnight or between 0200 and 0300. The strong winds usually blow for about half an hour, varying between five minutes and two hours.

Opinions differ about their origin. Some line-squalls originate in convection among the Sumatran ranges before sunset, and later travel eastward during the night, because they have been encountered in the evening on the east coast of Sumatra. Most probably line-squalls of this type form in the eastward movement of an air-stream boundary.

Radar observations over the Southern Malacca Straits during 1948 showed that, although cumuliform could develop right across the Straits at any time of night, it was mostly confined to the eastern side of the Straits. Captain A. Denny of Sepang (in Selangor near Port Dickson) describes an occasion when, although the sky was clear to the west at 2100, a vigorous 'Sumatra' crossed the Malayan coast two hours later. Aircraft reports have shown that thunderstorm activity by night is greatest in the eastern Straits, the region of maximum activity being very close to the Malayan coast a little before sunrise. It is likely, therefore, that most true 'Sumatras', as distinct from air-stream boundaries, form in the eastern Straits.

What factors contribute to the formation and movement of 'Sumatras'? The air over the region is conditionally unstable, and the distribution of temperature with height shows little variation from day to day. Throughout the season of the Southwest Monsoon, the monsoon westerlies or southwesterlies generally overlie the southerlies of the Trades, and no variations of this structure have been observed on nights when 'Sumatras' have occurred.

Practically every time a 'Sumatra' forms in the Straits, land breezes have preceded it along the Malayan coast. On other nights

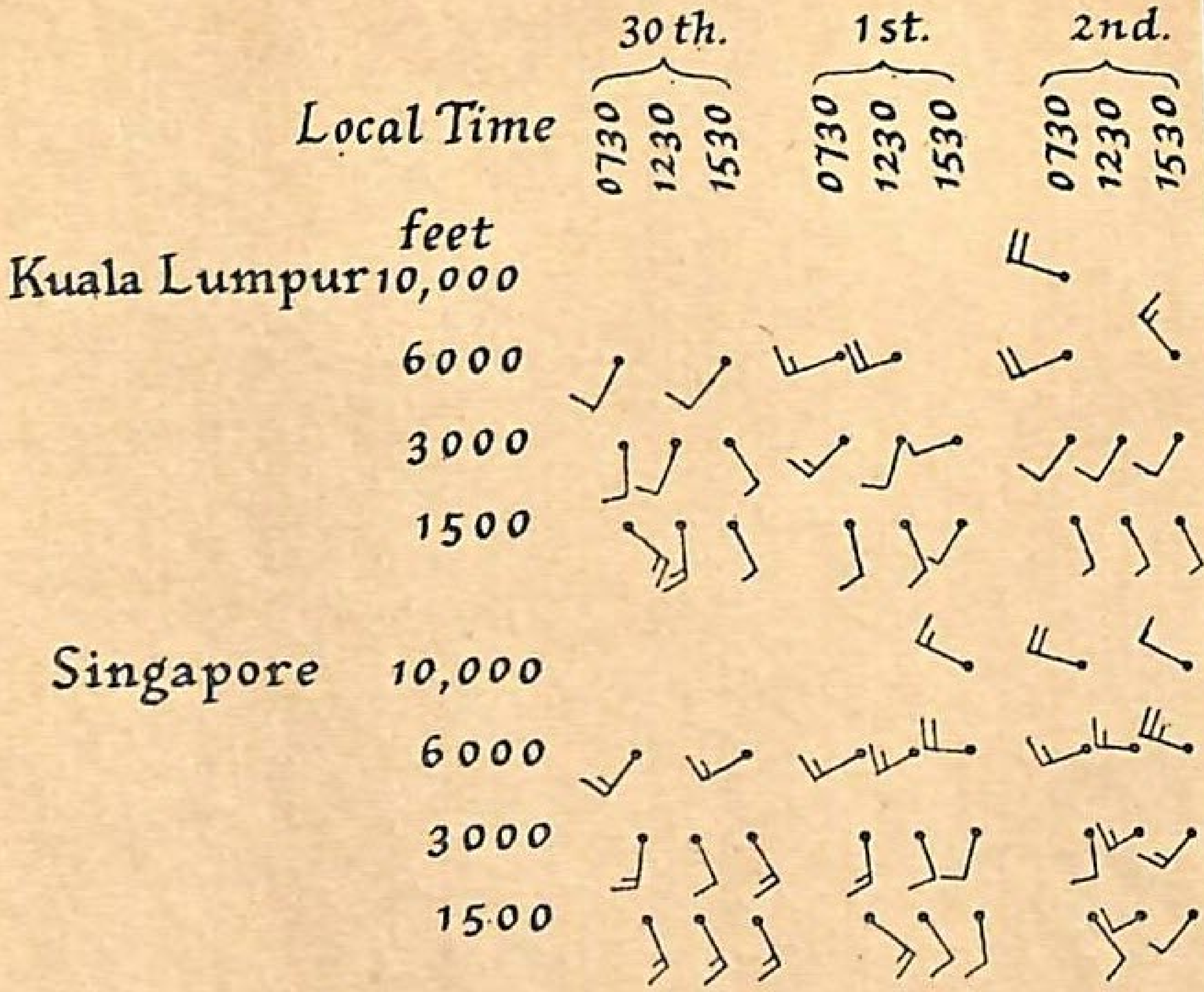


FIG. 105 Upper Winds, 30.6.36–2.7.36

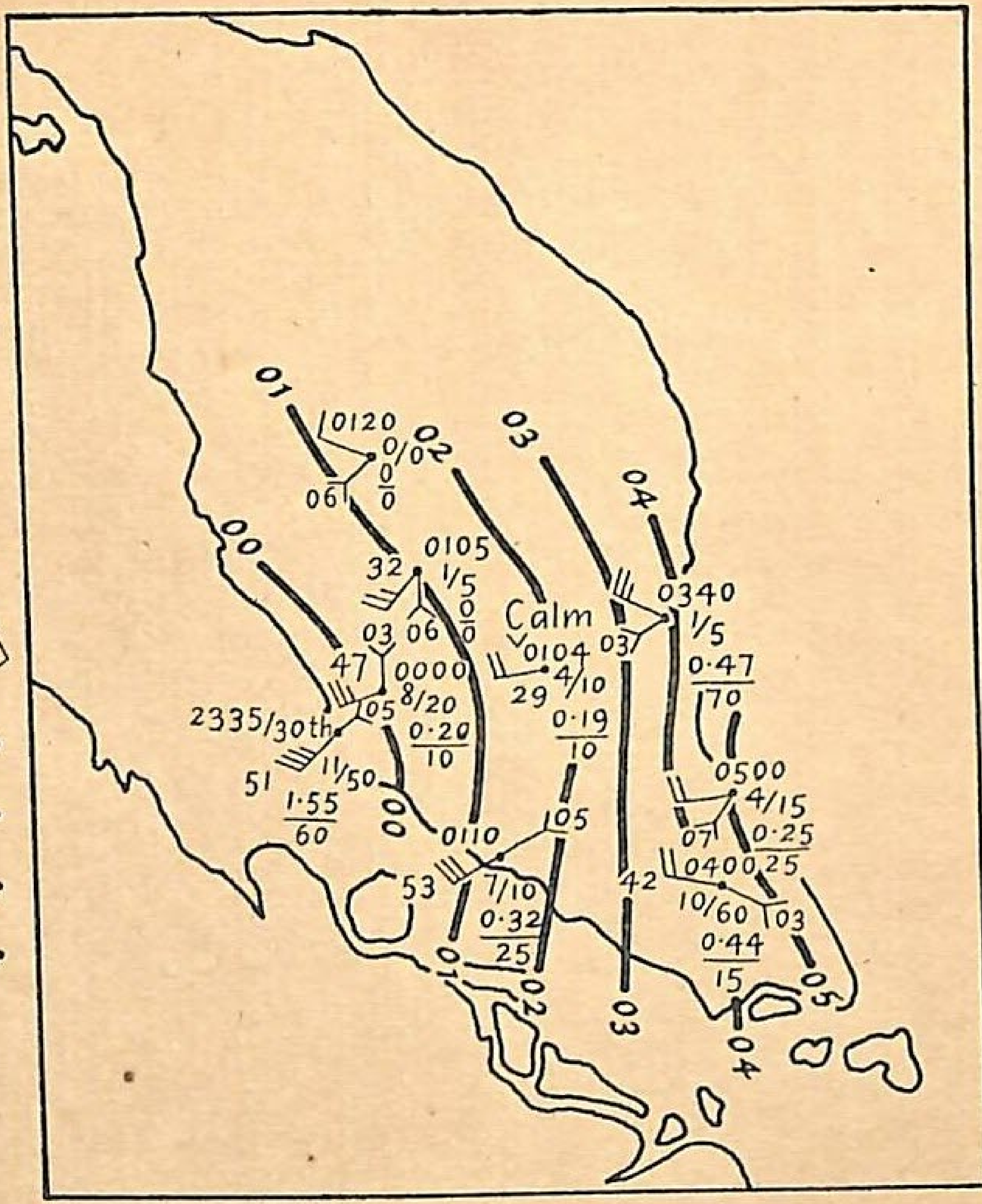


FIG. 106 Squall of 1.7.36

when the line-squalls might be ascribed to moving boundaries, the land breezes were not always established before the squall (cf. Figs. 104 and 106), making it probable that, on nights when 'Sumatras' occur, uplift is initiated by the land breeze undercutting the conditionally unstable air over the Straits, and that radiation from the tops of cumuliform columns further increases the instability and convection. Land breezes from Sumatra may also cause undercutting, but no evidence exists. Supporting the theory that the Malayan land breeze initiates the squall is the fact that clear evenings (which favour nocturnal radiation and early land breezes) precede most 'Sumatras'.

Figs. 105 and 106 show that the direction in which a 'Sumatra' travels bears no relation to the general wind structure at low levels, and it can also be shown that their occurrence and movement in no way depend on changes in the pressure distribution. Since the direction of its movement is the same as the wind direction at 6000 ft. (Figs. 105 and 106), this wind probably controls it. If the line of cumulonimbus is advanced bodily at 6000 ft., at lower levels clouds might be constantly dissipating to the west and re-forming to the east, a conception justifiable on the ground that the low-level southerlies are undiverted either before or after.

An alternative theory is that down-draughts from the cumulonimbus pass eastward with the 6000-ft. current, and undercut at ground-level to form another series of cumulonimbus columns down-wind. These down-currents, of small horizontal extent and short duration would not necessarily show in upper-wind observations, while the strength of the squall wind could be accounted for by convergence as the down-draught impinges on the ground.

'Sumatras' are mostly confined to the coast between Port Swettenham and Singapore, which might be explained by its curvature. From the southernmost tip of Malaya the coast runs nearly straight northwestward to Bukit Jeram, a section where the Malayan and Sumatran coasts are parallel. North of Bukit Jeram the Malayan coast curves in a great convexity until towards Penang another straight stretch runs to the north-northwest, while the Sumatran coast falls away to the northwest and the Straits widen appreciably.

The greatest frequency of 'Sumatras' is in the south where, although the coast is not concave, there is no likelihood of divergence in the lower winds. Divergence might be expected north of Bukit Jeram, where the Straits widen and the general orientation of airflow up to 3000 ft. is from the south or southeast. In the central section where the coast is convex, divergence would be greatest.

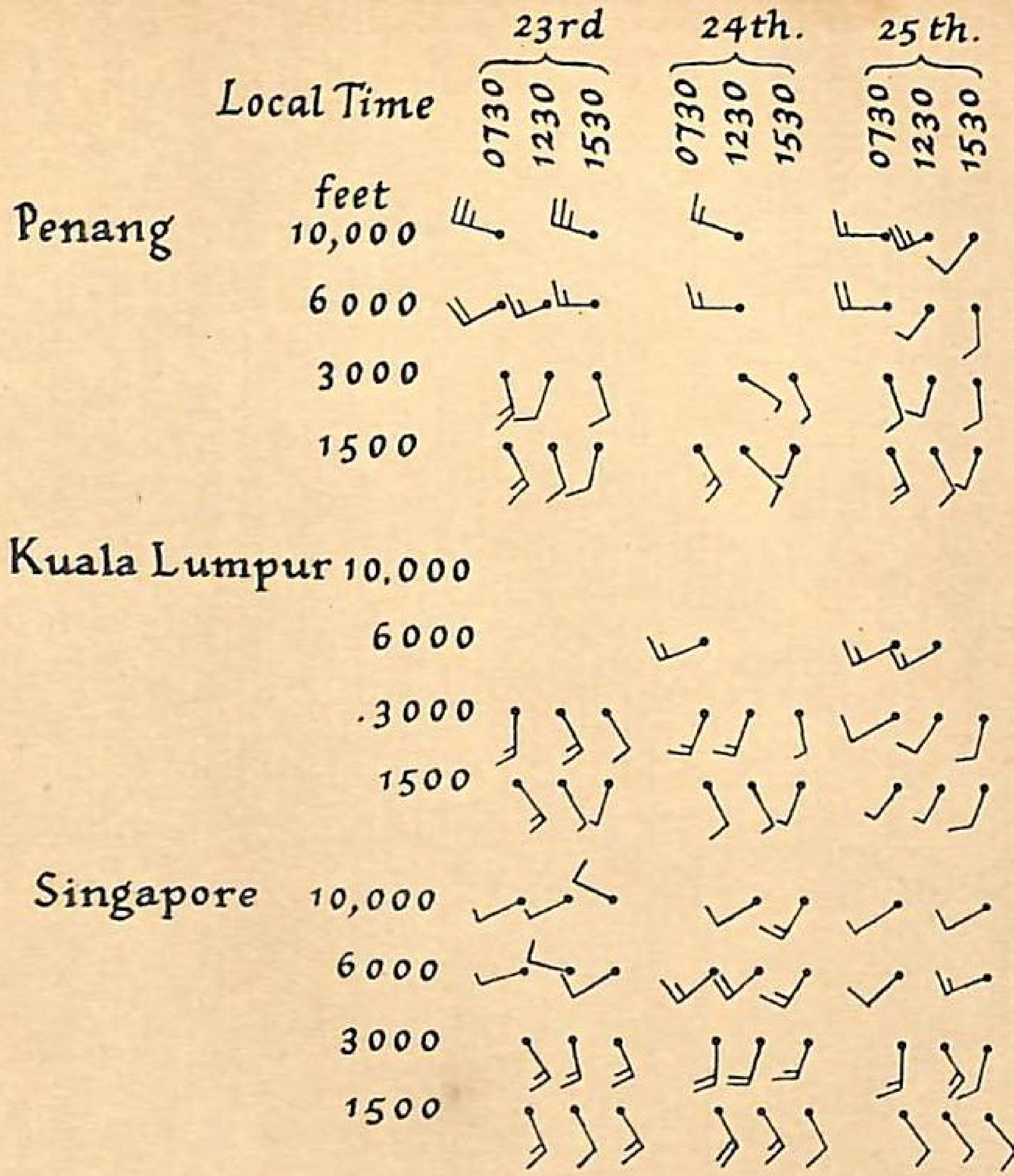


FIG. 107 Upper Winds, 23–25.7.37

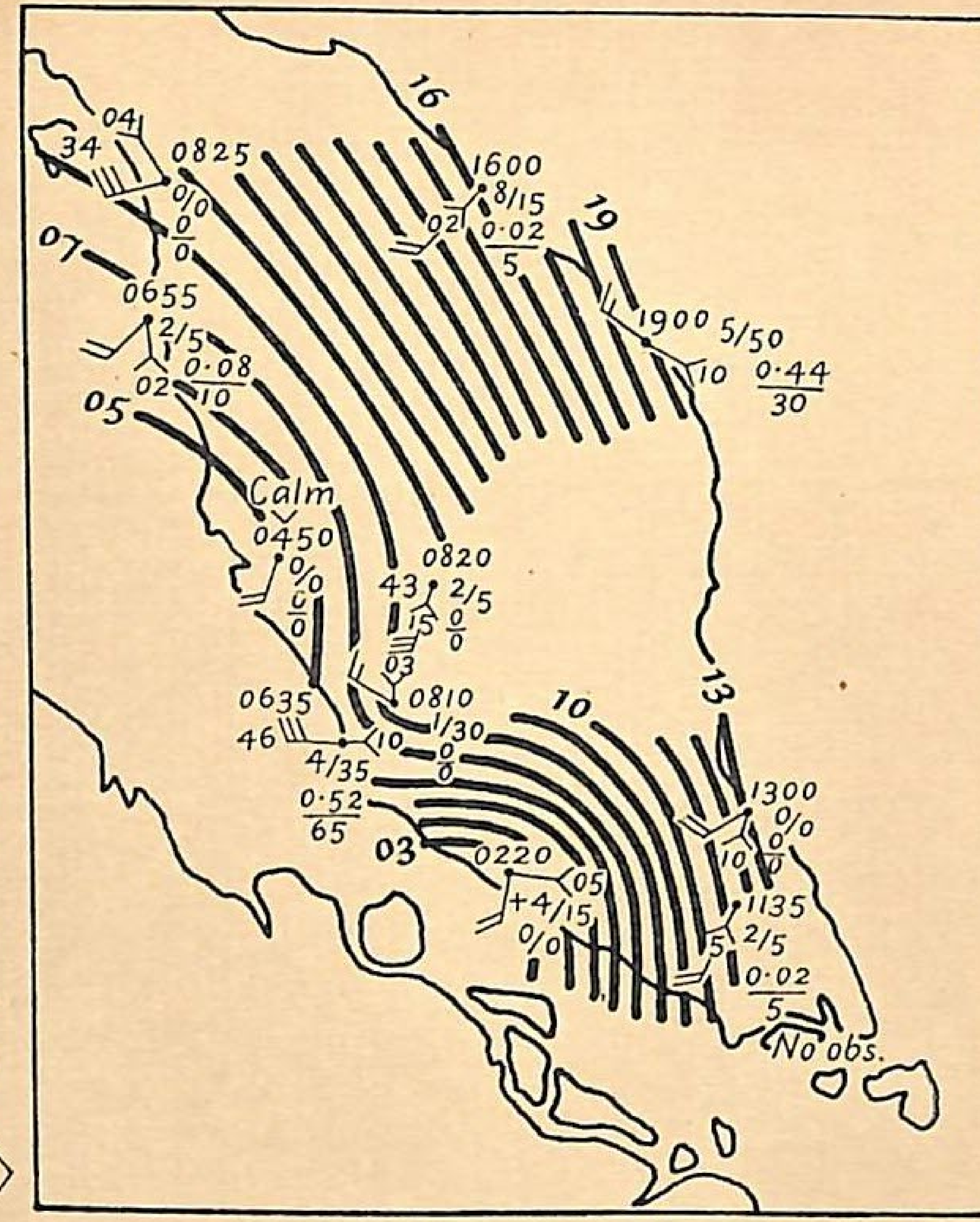


FIG. 108 Squall of 24.7.37

This reasoning is justified by a study of the early morning precipitation along the coast (Fig. 35), whose frequency is influenced by the convective activity brought about by convergence in the land breeze. Highest frequencies from May to September are found at Malacca and Bayan Lepas (Penang), but at Port Swettenham and Sitiawan in the convex zone the frequencies are low.

The tendency also shows in the early morning squalls from May to September (Fig. 100), whose frequencies are greatest at Malacca, negligible at Sitiawan and increasing again at Bayan Lepas. At Sitiawan, on the convex portion of the coast-line, every significant squall during the Southwest Monsoon is associated with a line-squall (Fig. 100), and reference to the original weather charts shows that on most occasions the squalls occurred during the eastward passage of a boundary. The region around Sitiawan is practically free from strong 'Sumatras'.

'Sumatras' tend to form in the north and south on the same night, as in Figs. 107 and 108. These are true 'Sumatras', with no discern-ible variation of upper winds as they pass. Small ground changes may occur at Sitiawan when the squalls to the south and to the north are large.

Other Squalls within a Stream

On page 163 was shown how the southeastward advance of the Southern Equatorial Boundary may be accompanied by a line-squall moving over Malaya. Similar line-squalls occur wholly within the air of the Southwest Monsoon when the Southern Equatorial Boundary is stationary, as in Fig. 110. The winds over Penang and Kuala Lumpur on the 15th October 1937 were north-westerly (Fig. 109) and in the Indian Southwest Monsoon. At Singapore, there were westerlies which may have been monsoon air from across the Sumatran ranges or diverted Southern Hemi-sphere Trades. The Southern Equatorial Boundary therefore lay either between Kuala Lumpur and Singapore or farther to the south.

The squall arrived on the coast in the early morning, probably initiated by a 'Sumatra', possibly by a disturbance which had travelled undetected over the Bay of Bengal. The evidence of ground and upper winds (Figs. 109 and 110) shows it unlikely that the squall reached Singapore. It is difficult to associate the vigorous squall at Mersing with the moving line-squall. A feature of this kind occurs regularly: when a line-squall appears within a stream and near the boundary, another squall or line-squall is often induced within the neighbouring stream so that the two, although converging on the boundary, have a velocity component along its length.

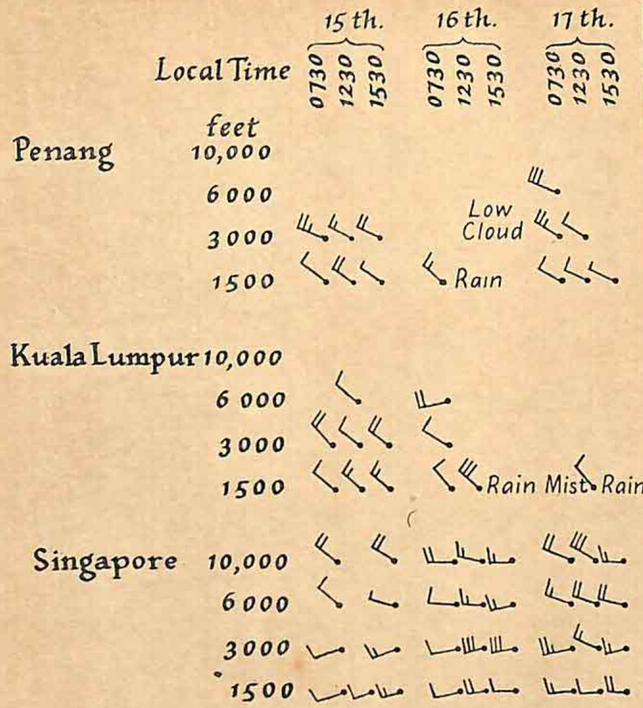


FIG. 109 Upper Winds, 15–17.10.37

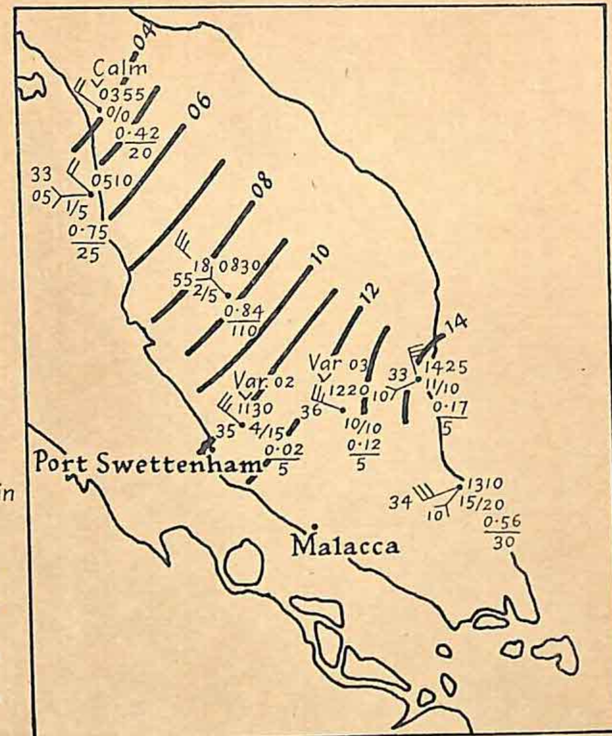


FIG. 110 Squall of 16.10.37

On 9th July 1937 (Fig. 111), the low-level flow at Penang showed that southerlies or southeasterlies, apparently Trades, covered Penang and Kuala Lumpur. Winds were more southwesterly at Singapore, but assuming that there were Trades in the north, then the southwesterlies in the south were also diverted Trades.

At Penang, the upper-level change to northwesterlies during the morning of the 10th represented an advance of the Southern Equatorial Boundary which did not reach Kuala Lumpur and Singapore until the 11th. While the boundary was advancing in the north, a separate squall wholly within the Trades crossed southern Malaya (Fig. 112). In the example, it might be considered that the more southerly squall was partly due to a 'Sumatra', but dual squalls of this nature may occur at any time of day.

When a squall occurs within a stream or when one stream advances displacing the boundary, a secondary line-squall is probably induced within the other stream so that the two travel parallel to the boundary.

5. The Local Problem

Line-squalls over Malaya conform to patterns: some are in movements of the air-stream boundaries and evident on the weather chart, while others consist of disturbances moving within a stream, not readily discernible on the chart and often originating from local causes.

Because relief influences the form and intensity of disturbances, the use of the synoptic chart in detailed study of the daily weather must be supplemented by considering local developments. These small-scale changes may be evident in observations made at a station at regular intervals, but because most squalls last only about half an hour, each station should report the passing of every 'significant squall'.

The problems of forecasting line-squalls are not peculiar to South-east Asia; they occur in many parts of the equatorial belt. More than one-fifth of the equatorial region is occupied by land or by islands adequate for creating a reasonable network of stations reporting at short intervals. Over the remaining four-fifths, islands are very scattered, so that not only is the synoptic reporting network unsatisfactorily spaced but it is impossible to supplement it with a squall-reporting network.

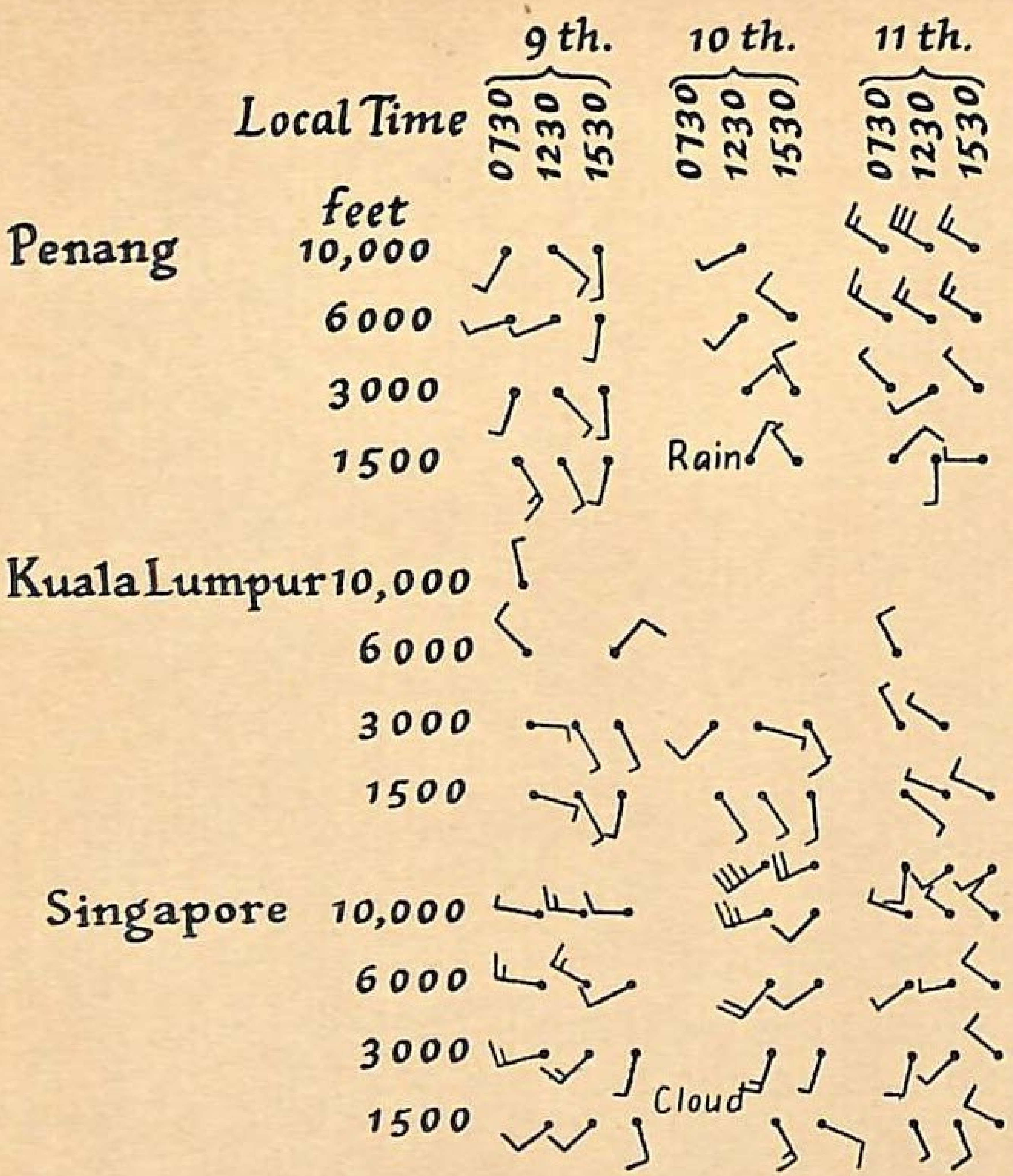


FIG. 111 Upper Winds, 9–11.7.37

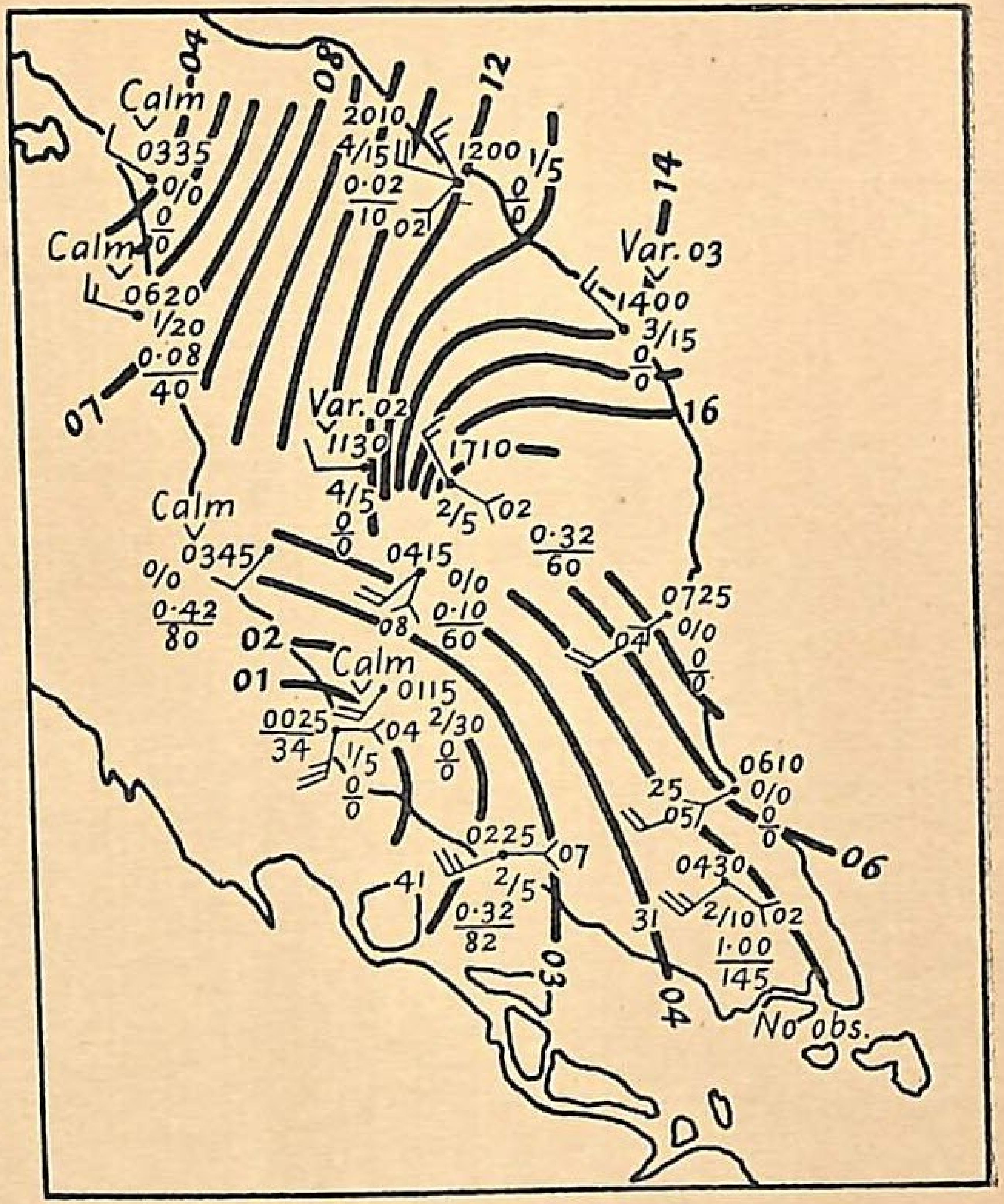


FIG. 112 Squall of 10.7.37

CHAPTER XIV

Forecasting Rain

Forecasts of rain are made by estimating probable movements of convergence zones and convergence lines by extrapolation from previous positions, allowance being made for changes in the wind-field which might influence the movements. When the lines or zones are slow-moving or stationary, predictions may be made by estimating the amount of horizontal convergence.

1. Measurement of Convergence

The distribution of horizontal convergence and divergence may be calculated from the observed winds. In this case 'convergence' includes not only zones within a stream but also convergence lines (i.e. air-stream boundaries accompanied by horizontal convergence). The method (after Forsdyke[44]) is given below and applied to conditions in Equatorial Southeast Asia.

(1) Break down each observed wind into the components '*u*' and '*v*' towards east and north respectively, and plot all the values of *u* and all those of *v* on two separate charts. Draw isopleths* of *u* and *v* on these charts. The observed winds at 5000 ft. at 0000 G.M.T., 2.1.51, are shown in Fig. 113. Owing to the great diversity in direction, it is not possible to draw stream-lines corresponding to these winds. The values of *u* and *v* and the isopleths therefrom are shown in Figs. 114 and 115 respectively, the units of *u* and *v* being chosen as knots to conform with international reporting procedure. The isopleths have been drawn at intervals of one knot.

(2) Select points on Figs. 114 and 115 where the isopleths show a definite gradient, and draw arrows in the direction of that gradient. The size of the gradient in knots per 50 miles may now be estimated from the intercepts of the isopleths on the arrows. These values are written beside the gradient arrows (thus <u>1·7</u>).

(3) Find the component of the gradient of *u* (at each of the points above) in the easterly direction and write the value beside

* Isopleths—generic term denoting lines joining points at which a particular element has the same value; or, more correctly, denoting lines drawn so as to separate areas recording values for a particular element above and below some chosen value.

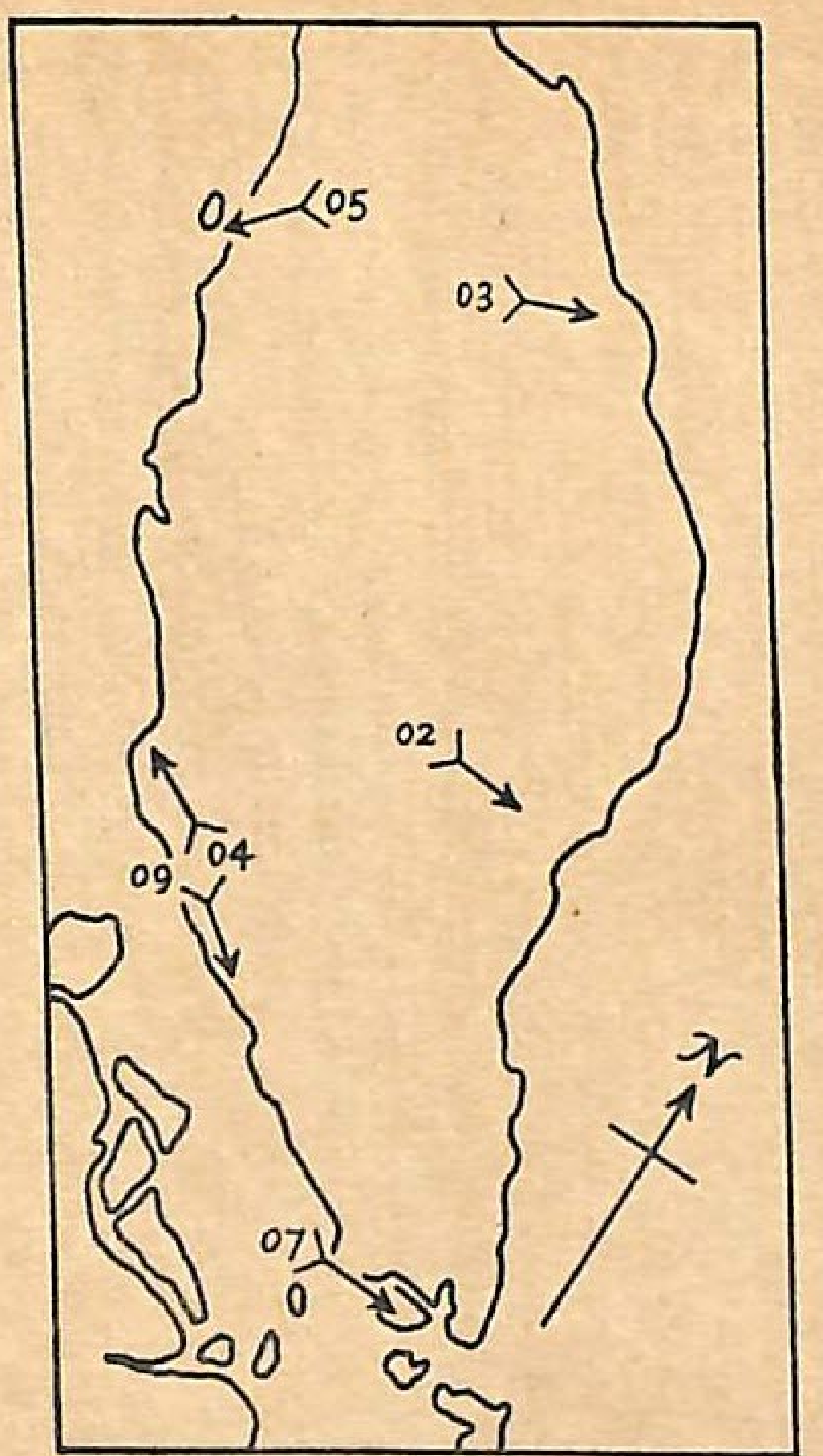


FIG. 113 Observed Winds at 5000 ft., 0000 G.M.T., 2.1.51 (Speed in knots)

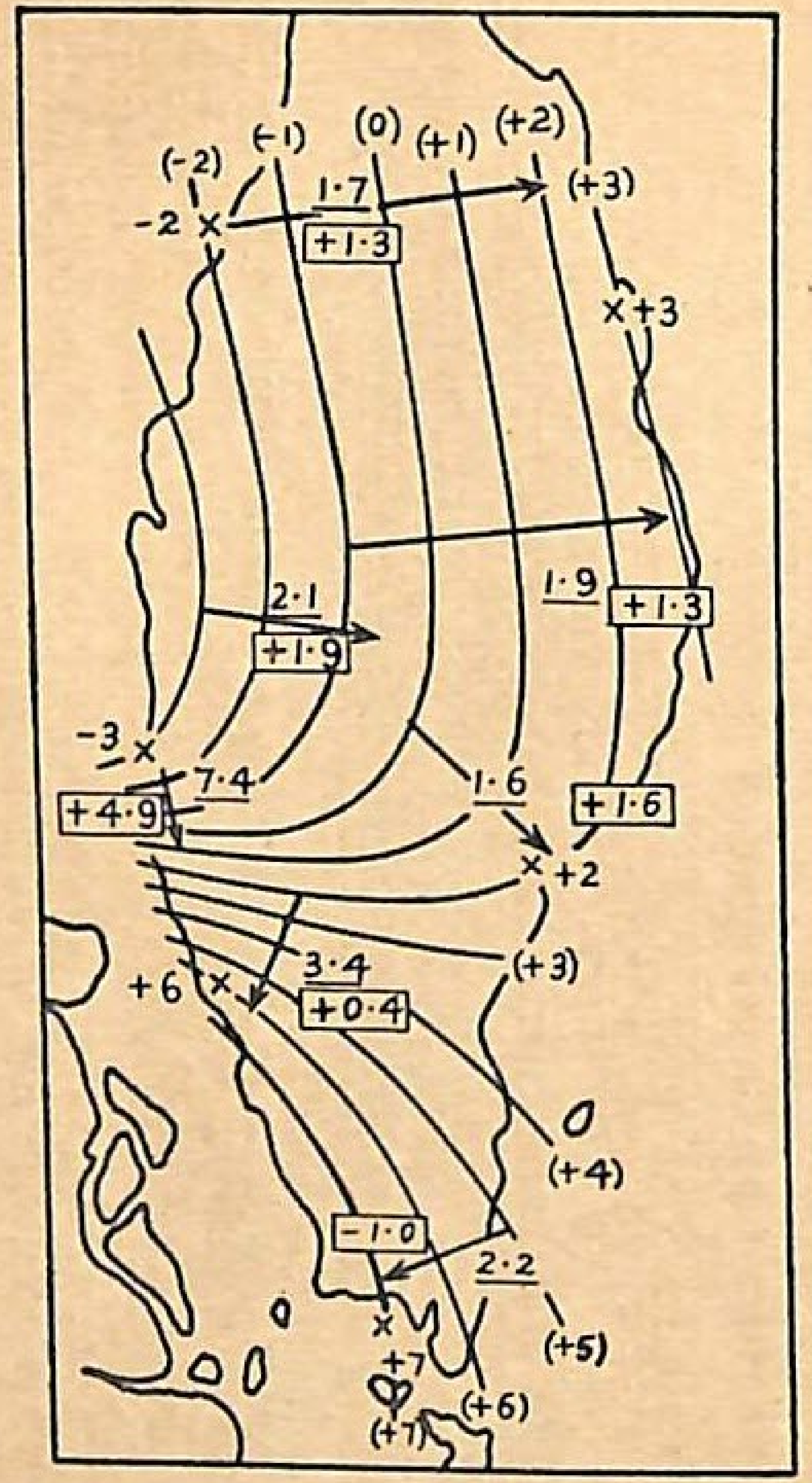


FIG. 114 Isopleths of u (in knots, 0000 G.M.T., 2.1.51 (Positions of stations shown by x)

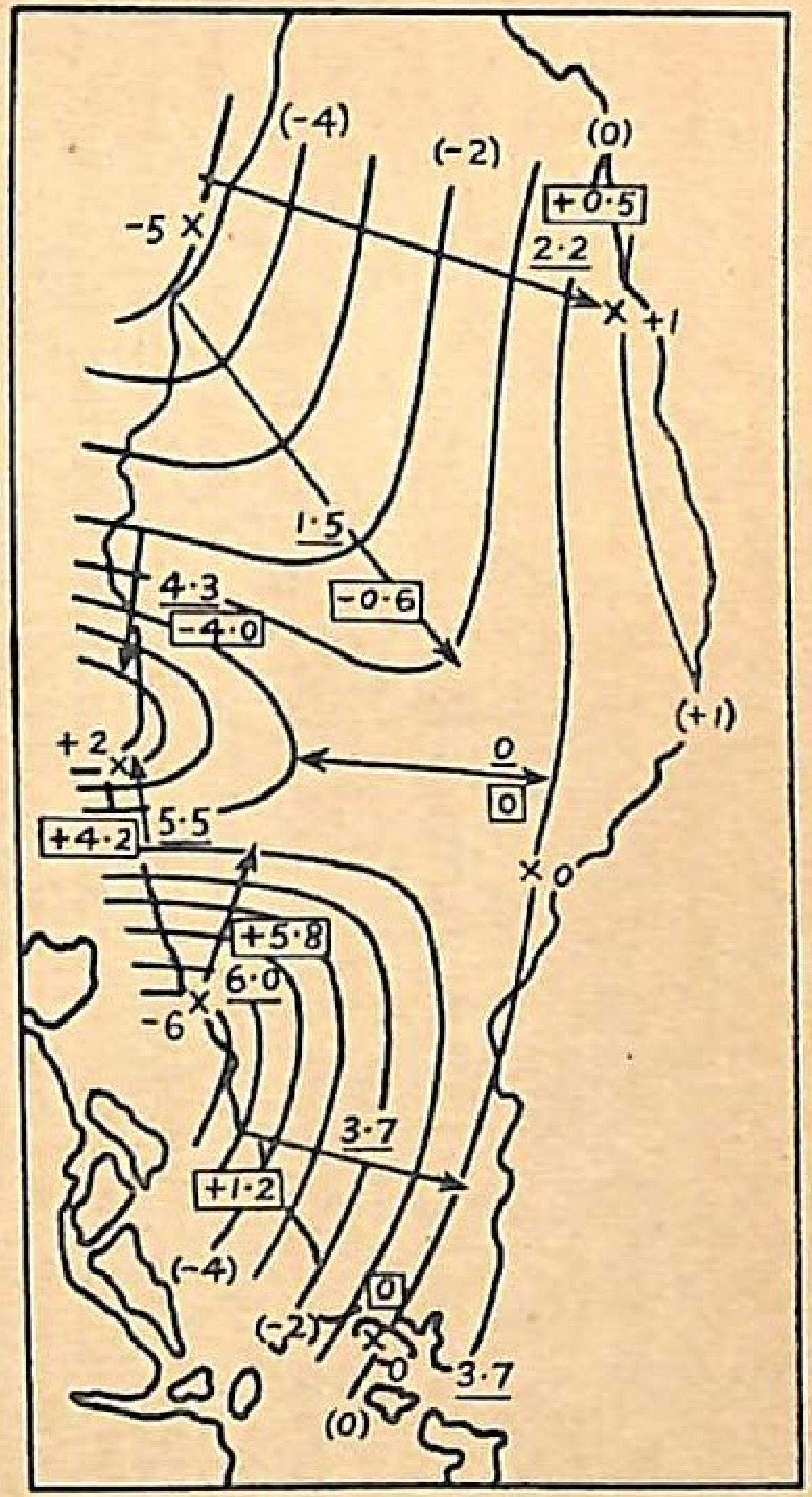


FIG. 115 Isopleths of v (in knots), 0000 G.M.T., 2.1.51 (Positions of stations shown by x)

the arrow in Fig. 114, showing whether the component is positive or negative (e.g. $\boxed{+1{\cdot}7}$). Determine in this way the components of the gradient of v in the northerly direction and plot on Fig. 115.

(4) On Fig. 116 plot each component of the gradient of u (to the east as found in (3)) at the positions occupied by the mid-points of the gradient arrows. Similarly, plot each component of the gradient of v (to the north) on Fig. 117. Draw isopleths on each of these charts.

(5) Trace the isopleths of Figs. 116 and 117 on to one chart, and write the algebraic sum of each pair of isopleths at every point of intersection. Then draw isopleths through these values (Fig. 118).

The final isopleths obtained by this process show the amount of divergence and convergence (in knots per 50 miles) over the region. Positive areas are those of divergence and are generally associated with fine weather; negative areas represent convergence, and large negative values are associated with regions of heavy rain. Isopleths of the field of divergence and convergence are determined thus at each of the reporting hours. From these charts rain is forecast when convergence is great or increasing with time, and fine weather is forecast when divergence is marked or increasing.

The type of weather may be correlated with the degree of divergence or convergence,[44] and the following table (after Kindle[68]) shows the relations (negative values referring to cases of convergence).

Approximate Weather	*Divergence—Knots per* 50 *Statute Miles*
Heavy Precipitation	-16
Slight Precipitation	$-1{\cdot}6 \times 10^{-1}$ to $-1{\cdot}6$
Very Little Weather	$-1{\cdot}6 \times 10^{-3}$
Very Light Subsidence	$+1{\cdot}6 \times 10^{-3}$
Moderate Subsidence	$+1{\cdot}6 \times 10^{-1}$ to $+1{\cdot}6$

Fig. 119 shows the amount of precipitation over Malaya for the 24-hour period beginning 0000 G.M.T. on 2.1.51. The features are:

(*a*) An area of moderate to heavy rain over Singapore and Southern Johore.

(*b*) The remainder of Malaya has experienced practically no rain except in the central ranges. One total of 1·32 inches was reported at Kajang, presumably a local thunderstorm because only 12 miles away in Kuala Lumpur there was no rain.

There is agreement between the degree of convergence and

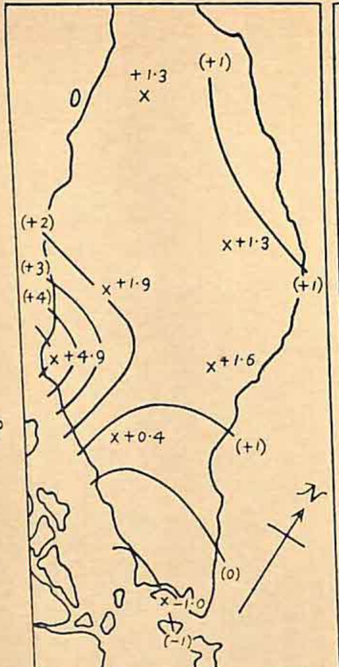


FIG. 116 Isopleths of Gradient of u, 0000 G.M.T., 2.1.51 (Positions of mid-points of gradient arrows of Fig. 114 marked by x)

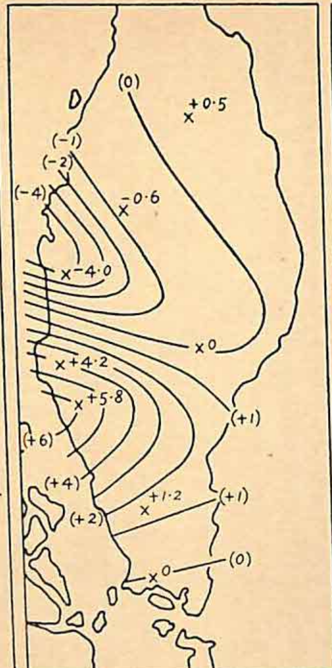


FIG. 117 Isopleths of Gradient of v, 0000 G.M.T., 2.1.51 (Positions of mid-points of gradient arrows of Fig. 115 marked by x)

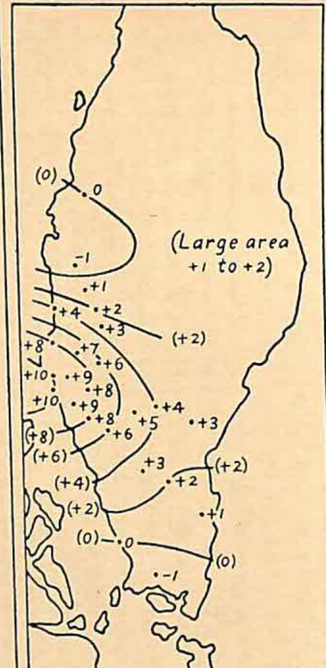


FIG. 118 Isopleths of Sum of the Gradients of u and v

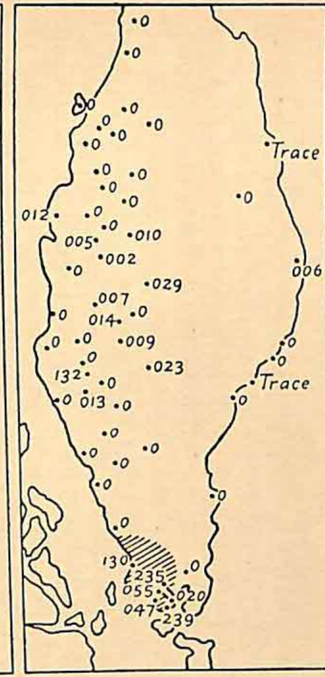


FIG. 119 Rainfall (in hundredths of an inch) from 0000 G.M.T., 2.1.51 to 0000 G.M.T., 3.1.51 (Positions of stations shown by dots; areas of rainfall greater than 0·50 in. shaded)

divergence shown in Fig. 118 and the distribution of rainfall in Fig. 119. Thus:

(*a*) The area about Sitiawan (4° N., 101° E.) was one of slight convergence and Kindle's table stipulates slight precipitation which probably did occur.

(*b*) Over the remainder of the country north of 2° N. there was general divergence, while the large degree of divergence in Southern Selangor (3½° N., 101½° E.) is equivalent to great subsidence.

(*c*) Fig. 118 shows convergence south of 2° N., but as the southernmost wind used in the computations was that of Singapore, it is hard to say what was its maximum value. Probably the centre of maximum convergence was south of Singapore, since Kallang (on the south of the island) reported a fall of 2·39 inches.

The areas of convergence and divergence were also determined for 0000 G.M.T. on 10.1.51 (Fig. 120) from the observed winds shown in Fig. 121. The features here are (*a*) divergence over the northeastern half of Malaya, and (*b*) convergence over southwestern Malaya and Singapore, with maximum convergence over Selangor. Reference to the map of rainfall for the period of 24 hours from 0000 G.M.T. on 10.1.51 (Fig. 122) indicates that the areas of maximum rainfall were Selangor and from Singapore to Southern Johore.

There is correlation between the distribution of rainfall and the areas of convergence, yet it is questionable whether consistently useful results may be obtained with the low density of upper-wind observations used in these examples. The convergence calculated in these examples is that within a narrow layer at 5000 ft., and to obtain a picture of total convergence through the atmosphere, similar calculations must be made at other levels and the results added.

In the examples given here, estimates of convergence and divergence near the coast are faulty because no wind reports over the sea are available. The upper winds at Kota Bharu, Penang and Kuantan are in themselves evidence of divergence over south-eastern Malaya on the 2nd January (Fig. 113). Similarly the slackening of wind from Kuantan to Port Swettenham and Malacca (Fig. 121) is sufficient evidence of convergence on the 10th January.

2. Reliability of Forecasts

We have stressed the importance of the contribution which may be made to the monthly rainfall by the sum of many isolated showers. Let us now examine their importance when forecasting daily weather. Suppose that the technique of forecasting is suffi-

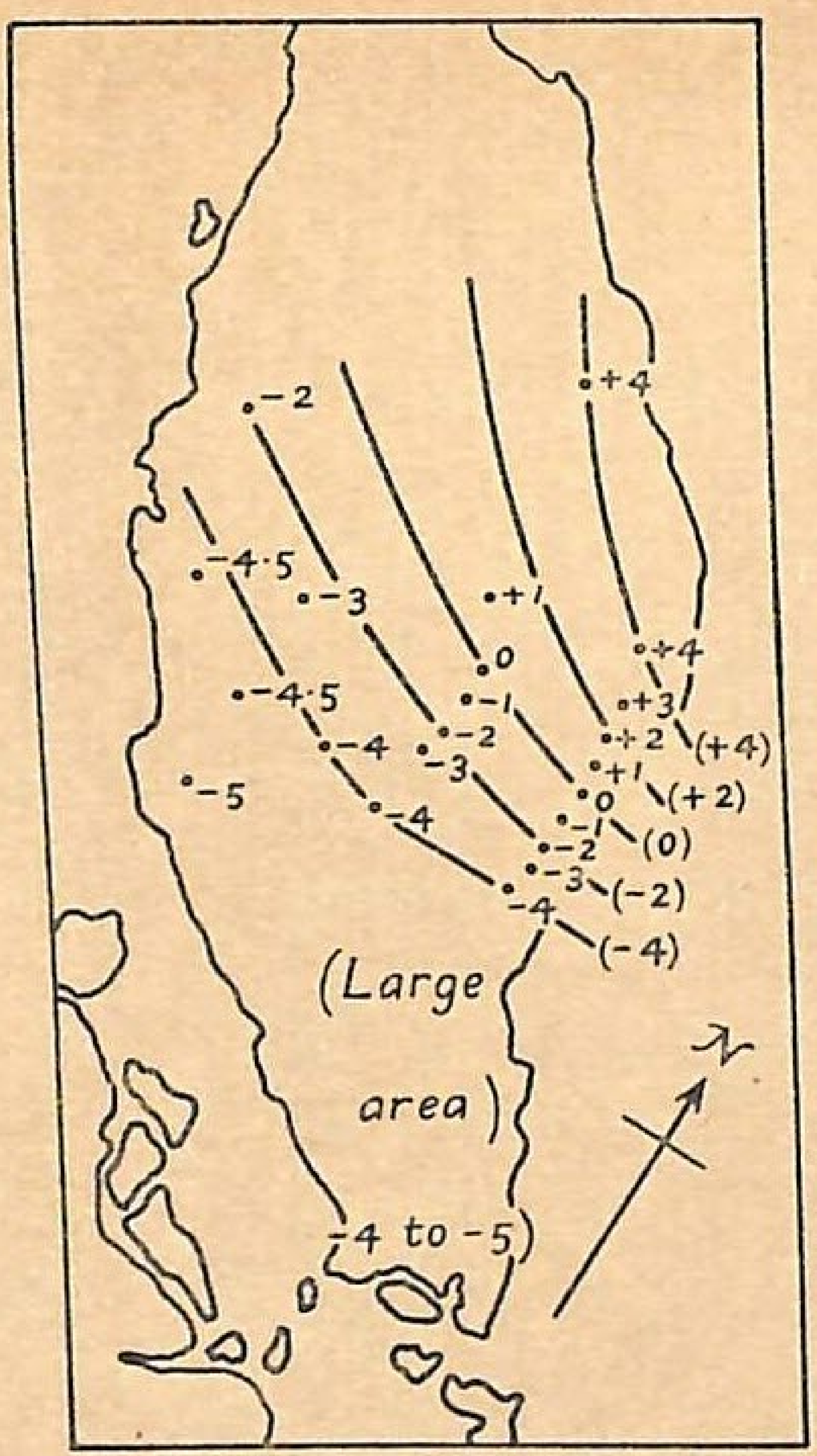


FIG. 120 Isopleths of Sum of Gradients of u and v at 5000 ft., 0000 G.M.T., 10.1.51

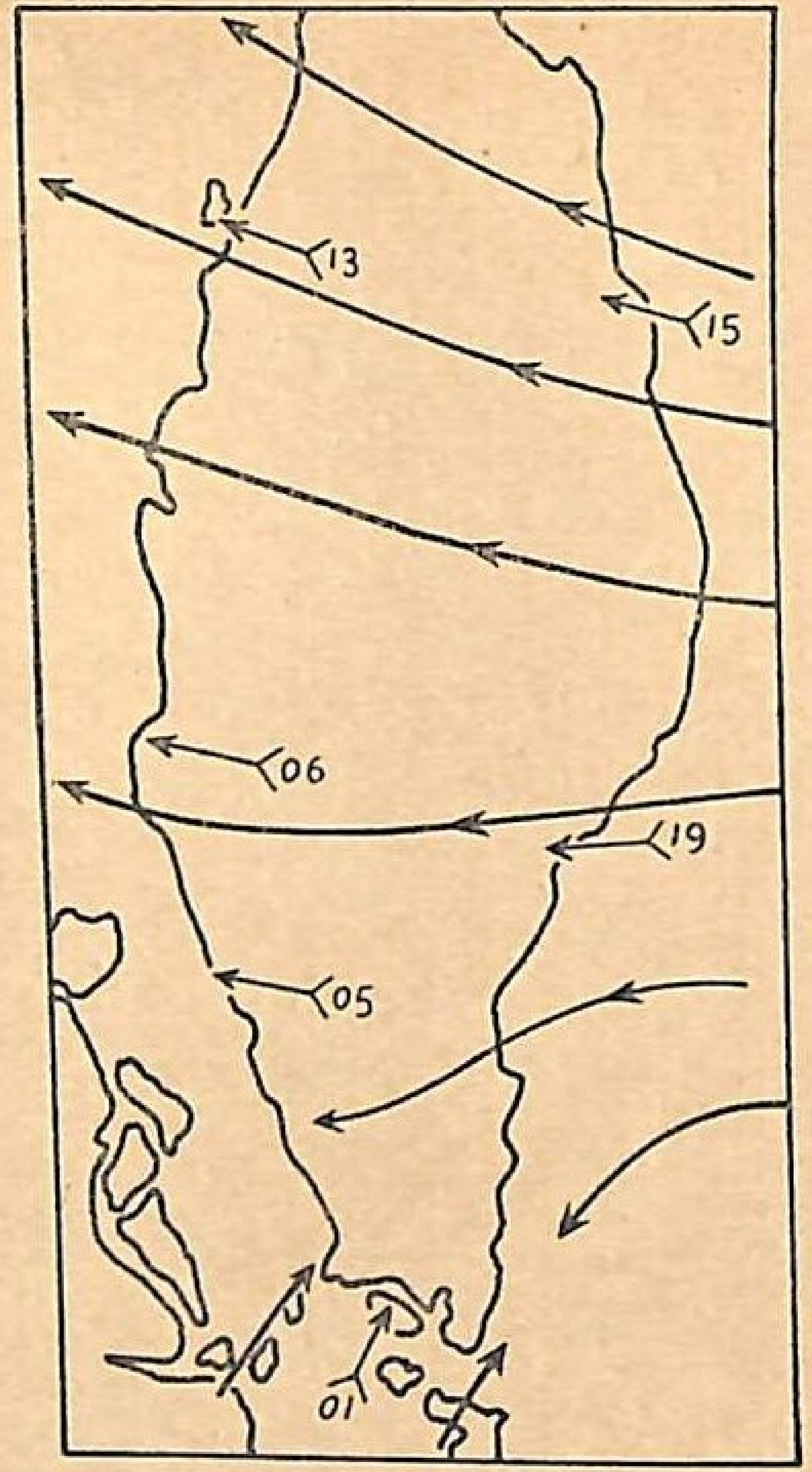


FIG. 121 Observed Winds (speed in knots) and Stream-lines at 5000 ft., 0000 G.M.T., 10.1.51

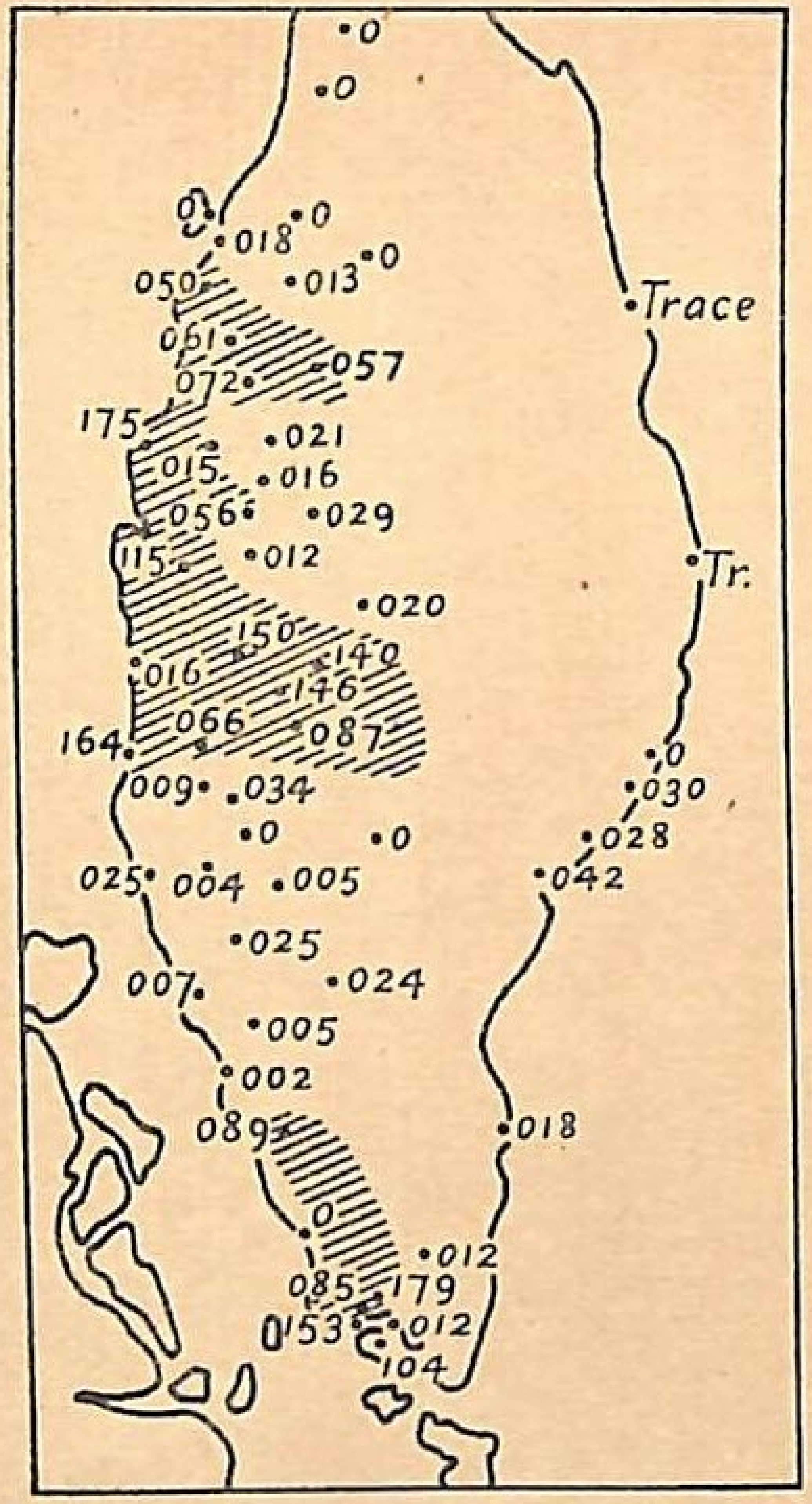


FIG. 122 Rainfall (in hundredths of an inch) from 0000 G.M.T., 10.1.51 to 0000 G.M.T., 11.1.51 (Positions of stations shown by dots; areas of rainfall greater than 0·50 in. shaded)

ciently well-developed to permit accurate prediction of wet days, showery days and fine days, and we will define these as follows:

(1) Fine days—when precipitation occurs over less than 25% of the forecast area.

(2) Showery days—when precipitation occurs over 25 to 65% of the forecast area.

(3) Wet days—when precipitation occurs over more than 65% of the area.

The minimum of precipitation considered in these cases is a fall of 0·10 inch in 24 hours.

An analysis has been made of conditions at Singapore* during the years 1948–50. Each day of the period has been classified as fine, showery or wet according to the above scale. The average number of days in each class was as follows:

Type of Day	*J.*	*F.*	*M.*	*A.*	*M.*	*J.*	*J.*	*A.*	*S.*	*O.*	*N.*	*D.*	*Year*
Fine .	14	12	14	11	11	19	19	19	19	14	10	9	171
Showery	8	11	8	11	11	5	5	8	7	9	8	12	103
Wet .	9	5	9	8	9	6	7	4	4	8	12	10	91

This means that, if the meteorologist's technique had permitted him to forecast correctly the 171 fine days, his forecasts would have been satisfactory for at least 75% of the area. Similarly, if he had predicted the 91 wet days, he would have satisfied the inhabitants of a considerable portion of the island (at least 65% of the total area).

There remain 103 showery days, which amount to nearly 30% of the year. Even supposing that the forecaster's technique permitted him to predict showers on these days, it must have been an unsatisfactory forecast to the man in the street, because only 25 to 65% of the total area actually encountered the showers. Another way of looking at this is that on 103 days each year the correct forecast would be for showery conditions, but the actual showers would only cover about one-fourth to two-thirds of the island. If we assume that the distribution of wet days at Singapore is typical, it is obvious that forecasting in the equatorial region must always be unsatisfactory during a great part of the year, although the forecaster may be capable of discriminating in his predictions between the wet, the dry and the showery days.

* The island of Singapore is approximately 24 by 13 miles. The percentage area of the island covered by precipitation has been computed in each case from the daily reports of six to eleven stations well-distributed over the island.

CHAPTER XV

Equatorial Air-stream Boundaries Elsewhere

The Equatorial Westerlies are confined to two regions. The first extends from the Central Indian Ocean to New Guinea, and the second covers a small expanse of the Eastern Pacific Ocean, Ecuador and Colombia. Elsewhere along the Equator the Trades meet in a single Equatorial Boundary, which at most times is easy to identify although greatly distorted. Away from the influence of the land masses, seasonal oscillations of the single boundary are not large and it is oriented roughly east–west. Near Africa and Australia, summer heating of the continents is so great that the development of monsoons broadens the seasonal oscillation by several degrees of latitude.

1. West Africa

In West Africa there exists one single boundary separating the Trades; those of the Northern Hemisphere are normally undiverted northeasterlies and, although those of the Southern Hemisphere are frequently turned to southwesterlies, the change is gradual and never sufficient for a second boundary to form at the diversion. The continent never becomes greatly cooled, and consequently the single boundary rarely passes south of 3° N. in the Gulf of Guinea, even during the southern summer. From the Guinea Coast northwards overland, there are two wind régimes—dry northeasterlies in the northern winter, and wet monsoon southwesterlies during the northern summer. Over land at the Equator, there are two periods of rain associated with movements of the boundary, while farther south there is a single small maximum of rainfall near the end of the southern summer.

During January and February, the surface Equatorial Boundary over the Atlantic lies along the Equator and close to the Guinea Coast (Fig. 124). It slowly advances northward as far as 8° N. during March and April. Early in the year, cooling in the Sahara maintains high pressures which may exceed 1022 millibars,[69] while those in the equatorial trough are about 1010 millibars. The large difference of pressure between the two areas causes a movement of

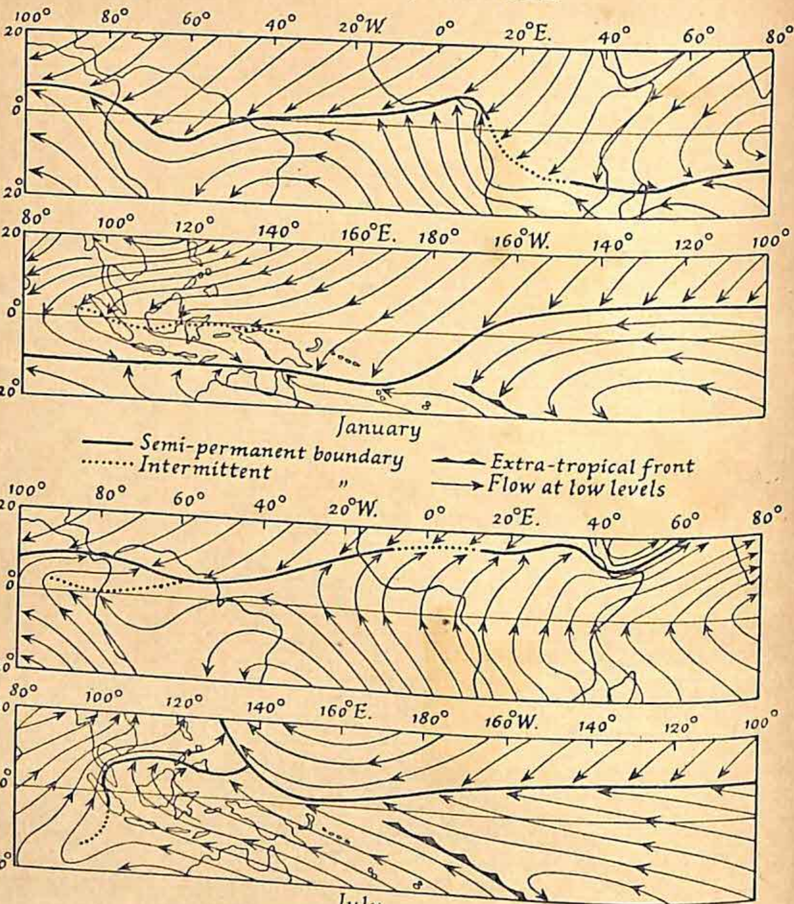


FIG. 123 Mean Positions of the Major Equatorial Air-stream Boundaries at Low Levels

air such that northeasterlies blow fairly consistently overland north of the boundary.

At the beginning of the year, southwesterly sea breezes frequently replace the northeasterlies on the Guinea Coast, yet they never reach higher than 4000 ft., above which the northeasterly is undisturbed. Above 10,000 ft. there are easterlies or northeasterlies throughout the whole year, so that the slope of the boundary is always upwards to the south.

During May and June the boundary moves farther northward with the collapse of the Sahara High, and the axis of equatorial low-

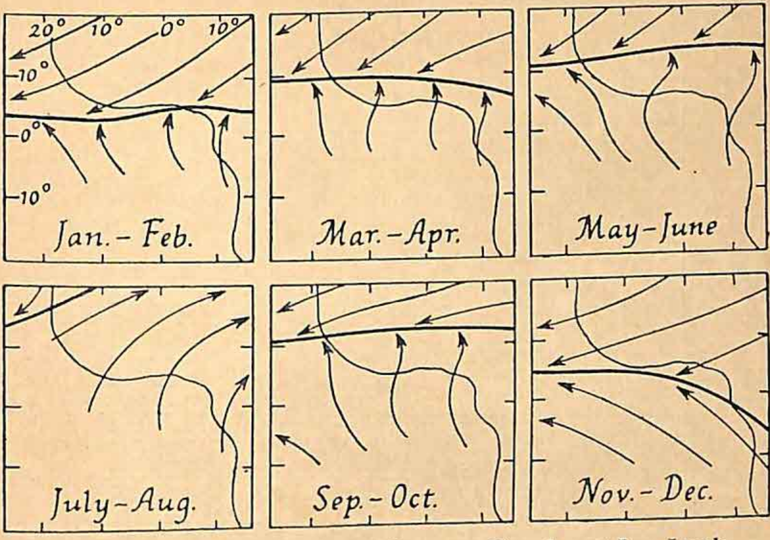


FIG. 124 West Africa: Equatorial Air-stream Boundary at Low Levels

pressure assumes a northeast–southwest orientation. As it advances northward, a broad stream of southwesterly winds follows to cover the whole region by July. This 'Southwest Monsoon' when well-established extends to about 10,000 ft. over the Guinea Coast, decreasing its altitude in the north.

From July to October, the boundary inland lies to the north of Nigeria and the Gold Coast, while from Senegal it curves west-southwestward across the Atlantic, reaching the American coast north of the Amazon.

The southward movement of the boundary is rapid during November and December, when its orientation in the Atlantic is nearly east–west along 4° N., but over the Gulf of Guinea it bends more to the southeast to cross the West African coast about 7° S. A renewal of the northeasterlies accompanies the movement of the boundary southwards.

The seasonal differences of weather in this region are associated with the differences in orientation of the main wind streams. Little rain occurs with the northeasterlies: there are frequent heavy falls during the Southwest Monsoon (Table 6). Mean temperatures are about 80° F., and though maxima are not very high at the coast, they frequently exceed 100° inland.

The northeasterly wind, often termed the 'Harmattan', is very dry and contains much dust or fine penetrating sand which reduces

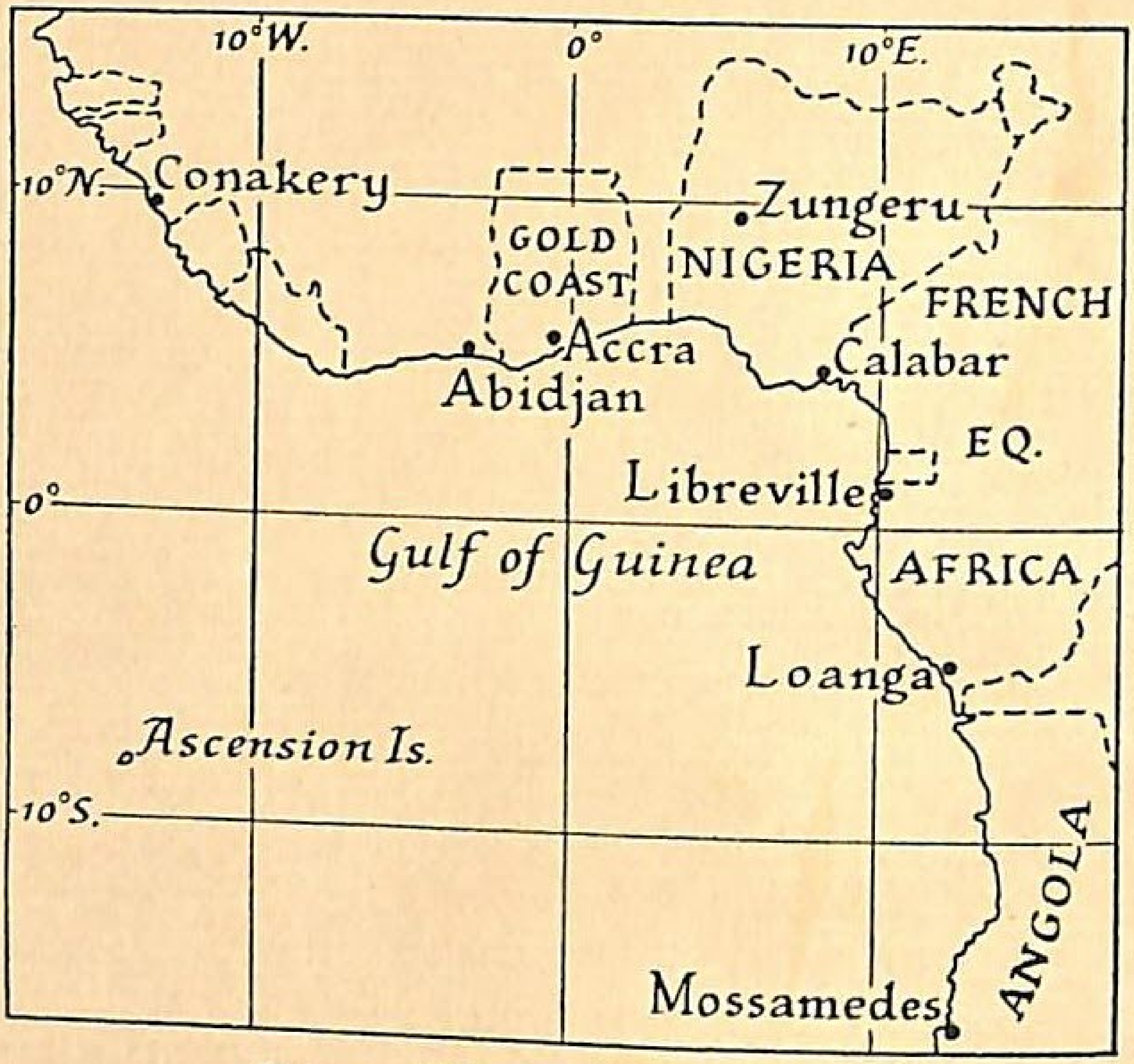


FIG. 125 West Africa: Key Map

visibility. Relative humidity is low on days when the Harmattan is blowing, and least when it is strongest.[70] The following table shows[70] the frequency of days of Harmattan at Zungeru (Northern Nigeria) in relation to the humidity.

TABLE 5

RELATIVE HUMIDITY DURING THE HARMATTAN

Month	*J.*	*F.*	*M.*	*A.*	*M.*	*J.*	*J.*	*A.*	*S.*	*O.*	*N.*	*D.*
No. of Days of Harmattan	22	21	10	1	0	0	0	0	0	6	24	20
Mean Relative Humidity when no Harmattan	46	52	57	67	71	80	84	84	82	78	69	48
Mean Relative Humidity when Harmattan Blowing	37	34	36	67	—	—	—	—	—	77	50	40

The rainfall of the Southwest Monsoon is greatest on the extreme west coast between 7° and 10° N. Inland there is a general decrease except in the mountains, and a similar decrease along the coast southwards (Table 6). Early morning fogs occur during the rains in coastal regions but rarely inland except in low-lying forest.

Violent thunderstorms with squalls may occur on the coast at or

Table 6

West Africa: Mean Monthly Rainfall in Inches

Place	*Lat.*	*J.*	*F.*	*M.*	*A.*	*M.*	*J.*	*J.*	*A.*	*S.*	*O.*	*N.*	*D.*	*Year*
Zungeru (inland) .	9° 48′ N.	0·0	0·0	0·5	2·4	4·8	6·7	7·5	9·0	10·8	3·5	0·1	0·1	45·4
Conakery (coastal) .	9° 30′ N.	<·1	0·1	0·3	0·9	6·2	22·0	51·1	41·5	26·9	14·6	4·8	0·3	168·9
Abidjan ,, .	5° 19′ N.	1·6	2·1	3·9	4·9	14·2	19·5	8·4	2·1	2·8	6·6	7·9	3·1	77·1
Accra ,, .	5° 33′ N.	0·6	1·1	1·9	3·7	5·2	7·0	1·6	0·6	1·2	2·3	1·3	0·9	27·4
Calabar ,, .	4° 58′ N.	2·1	2·7	6·4	7·9	11·9	15·7	16·9	16·2	16·3	12·8	7·5	2·1	118·5
Libreville ,, .	0° 23′ N.	9·8	9·3	13·2	13·4	9·6	0·5	<·1	0·7	4·1	13·6	14·7	9·8	98·9
Loanga ,, .	4° 39′ S.	5·4	6·7	6·4	8·0	3·9	<·1	<·1	<·1	0·4	4·1	6·6	6·6	48·1
Ascension Is. .	7° 56′ S.	0·1	0·1	0·6	1·7	0·3	0·6	0·3	0·3	0·3	0·2	0·1	0·1	4·6
Mossamedes (coastal) .	15° 12′ S.	0·1	0·2	0·7	0·7	0·0	0·0	0·0	0·0	0·0	<·1	<·1	<·1	1·9

Table 7

East Africa: Mean Monthly Rainfall in Inches

Station	*Lat.*	*J.*	*F.*	*M.*	*A.*	*M.*	*J.*	*J.*	*A.*	*S.*	*O.*	*N.*	*D.*	*Year*
Berbera (British Somaliland) . . .	10° 22′ N.	0·12	0·31	0·73	0·49	0·39	0·00	0·11	0·05	0·03	0·05	0·03	0·07	2·38
Entebbe (Uganda) .	0° 4′ N.	2·57	3·57	6·27	10·08	9·60	4·77	3·00	2·95	2·95	3·74	5·17	4·57	59·24
Nairobi (Kenya) . .	1° 16′ S.	2·01	2·81	4·44	7·95	6·55	1·65	0·65	1·03	1·65	2·75	4·34	3·57	39·40
Dar-es-Salaam (Tanganyika) . . .	6° 50′ S.	2·76	3·21	5·63	11·83	7·44	1·11	1·14	1·06	1·38	2·25	2·75	3·12	43·68
Antananarivo (Madagascar) . . .	18° 55′ S.	12·81	10·52	7·02	2·56	0·90	0·42	0·30	0·31	0·81	1·21	6·09	11·40	54·35

before the start of the Southwest Monsoon and these, when well-developed, are locally termed 'tornadoes'.

2. The Sudan

The effect of the Southwest Monsoon extends far into Central Africa, and its influence is considerable even at Khartoum ($15\frac{1}{2}°$ N.) as in the following table.[71]

TABLE 8

FREQUENCY OF WINDS FROM N.-NE. AND S.-SW. AT KHARTOUM AND MEAN MONTHLY RAINFALLS

Month	*J.*	*F.*	*M.*	*A.*	*M.*	*J.*	*J.*	*A.*	*S.*	*O.*	*N.*	*D.*
Percentage of Winds from N.-NE.	93	92	87	72	47	11	2	3	9	45	83	94
,, ,, ,, ,, S.-SW.	0	0	0	3	17	49	74	71	48	15	1	0
,, from Other Directions	7	8	13	25	36	40	24	26	43	40	16	6
Rainfall (in inches) . .	Tr.	Tr.	·04	Tr.	·12	·31	2·17	2·64	·63	·16	Tr.	0

According to El-Fandy,[72] the mean position of the Equatorial Boundary (termed by him the 'Intertropical Convergence Zone') during the northern summer is along lat. 15° N. over the Western Sudan. Farther east it lies at about 18° N. between Khartoum and Port Sudan, thence curving towards the southeast into the Gulf of Aden (Fig. 126). Over the Sudan, as in West Africa, the boundary slopes upwards to the south, and at 10,000 ft. a current (originally

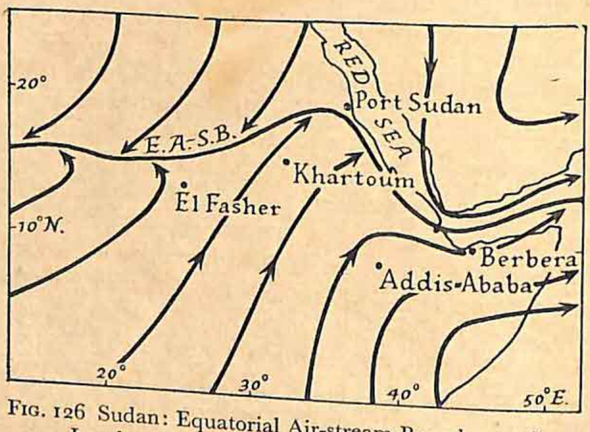


FIG. 126 Sudan: Equatorial Air-stream Boundary at Low Levels—Northern Summer (After El-Fandy)

a northeasterly stream over Arabia and the Red Sea) sweeps across the Sudan as an easterly flow. The ground boundary is not normally active, and disturbed weather mostly occurs within the southerlies or southwesterlies hundreds of miles from it.[73]

Over the Sudan and on the West African coast, the Southwest Monsoon is the rainy season (Table 8). During the period April–July in the Northern Sudan, occasional severe dust storms termed 'Haboobs' occur. In such a storm, columns of sand often extend to 3000 ft. or higher, while it is accompanied by gusts to 30 or 40 m.p.h. and, later in the season, by thunderstorms. It originates largely as a result of excessive heating of air over a large area for four to five days prior to its advent, leading to extreme instability in the lowest 10,000 ft.[74] The passage of a 'Haboob' is often followed by a marked temperature drop.

3. East Africa and the Indian Ocean

In January and February, the mean position of the boundary separating the Trades of both hemispheres is across Northern Mozambique and the northern tip of Madagascar (Fig. 127*a*). East of Madagascar it curves towards the east–northeast and follows 12° S. over the Southern Indian Ocean. Its position early in the year is across the Timor Sea, often encroaching on the Australian coast between Melville Island and Cape Leveque when the Northern Hemisphere Trades, diverted to northerlies or northwesterlies, advance over Arnheim Land as a summer monsoon.

At times between December and mid-March, the surface boundary over East Africa comes as far as 22° S., and frequently enough to influence the Madagascar rainfall. Thus at Antananarivo (19° S.), 75% of the annual rainfall occurs during these four months (Table 7).

During the southern summer, the Southeast Trades undergo little diversion other than that caused in the lowest layer by relief, and orientation over the ocean is mainly from east-southeast. The Trades of the Northern Hemisphere, on the other hand, are diverted to northerly or northwesterly about 10° S. (Fig. 127*a*), yet never turn sharply to form a distinct second boundary.

The Trades of the Northern Hemisphere nearly always overlie those of the Southern Hemisphere, and the slope is between 1 in 200 and 1 in 300, often persisting to 20,000 ft. above the ground.[59]

From March onwards, the boundary moves rapidly northward, and a general increase of rainfall extends as far as 5° N. Not all this increase may be attributed to convergence on the boundary, because there is increased convection inland associated with a

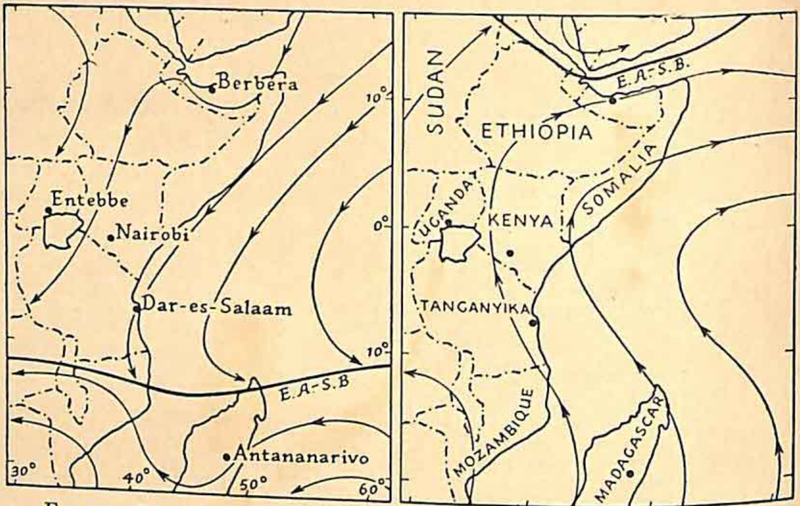


FIG. 127 East Africa: Equatorial Air-stream Boundaries at Low Levels
(*a*) January (*b*) July

decrease of winds before it passes. The coast is exposed to the Southeast Trades after the boundary passes, so that relief combines with convection to prolong the rainy season.

The boundary moves north of 5° N. by the end of April, associated with negligible rain, and, as the Southern Hemisphere Trades become diverted to a more westerly orientation, a föhn effect keeps rainfall low in Somaliland. The westerlies or 'Turned Trades' frequently flow completely across the Indian Ocean to contribute to the Equatorial Westerlies of Southeast Asia, the curvature of the stream being too slight to form a second boundary on the East African coast.

In May, the boundary lies along 7° or 8° N. over Somaliland and along 10° or 12° N. farther to the east, where it is the Northern Equatorial Boundary marking the onset of the Indian Southwest Monsoon.

There is little further advance of the boundary in the west from June to September, and it takes up a position along the Gulf of Aden. It curves sharply to the northeast because the Indian Southwest Monsoon has then completely covered India and Southeast Asia. The Trades from the Southern Hemisphere then have a true westerly orientation north of the Equator. At this time of year when westerlies are running side by side with the undiverted Southeast Trades over the Eastern Indian Ocean, conditions favour the formation of the Southern Equatorial Boundary.

Towards the end of October, the boundary passes south over Somaliland again, with little convergence until it nears the Equator in November. Increased rainfall then marks its passage, illustrated by the secondary maxima of rainfall at Entebbe and Nairobi (Table 7). A similar secondary maximum occurs at Dar-es-Salaam (7° S.) during December, though less pronounced because the boundary is then moving rapidly southward when the Northern Hemisphere Trades become reinforced by a northeasterly monsoon flow across the Arabian Sea from western India.

It is difficult to describe the seasonal changes over East Africa wholly in terms of the boundary and an alternation of the Trades. Away from the flat coastal belt, the country rises steeply and the climate falls into several regional types related to altitude. Convectional and relief effects are so great in some districts that changes due to the passage of the boundary may be masked.

4. The Western Pacific

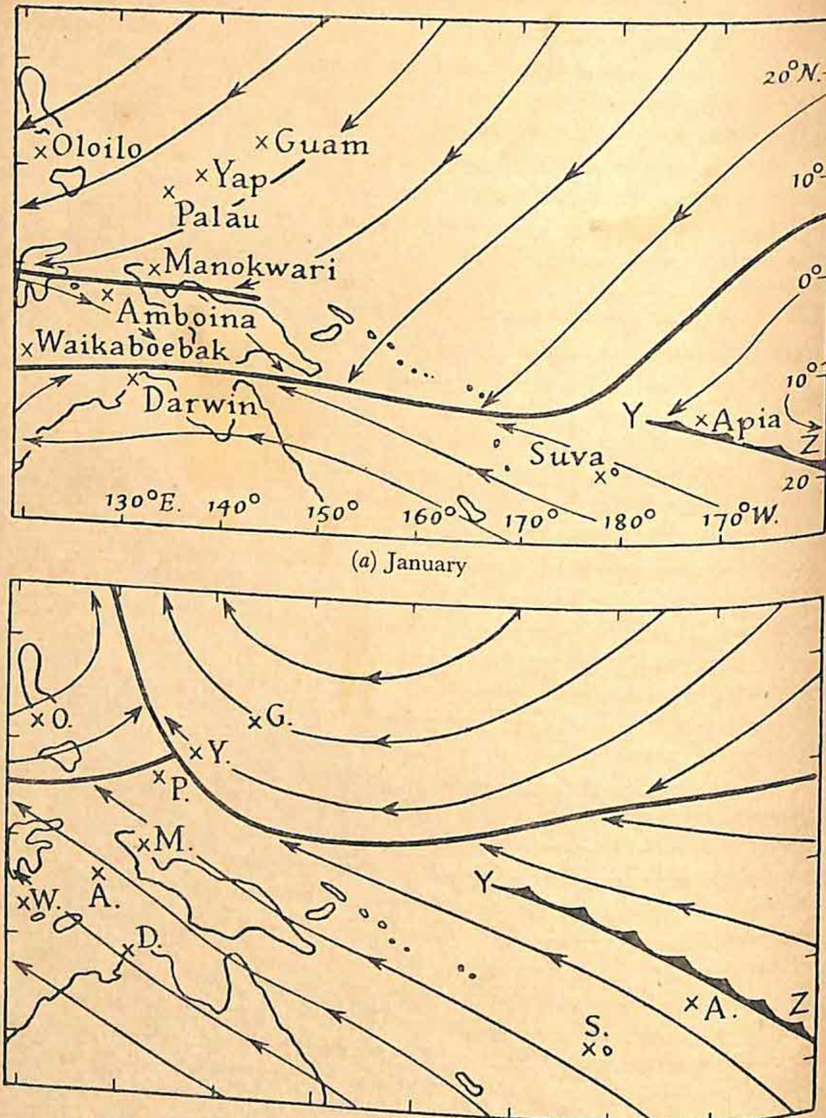
The separate existence of Northern and Southern Equatorial Boundaries may be traced east of Southeast Asia.

During January, the Southern Boundary lies along 10° or 12° S. north of Australia, while the Northern Boundary, intermittent in appearance, is close to the Equator (Fig. 128*a*). Between these boundaries are northwesterlies or westerlies diverted from the Northeast Trades, but east of 150° E. the Northern Boundary never forms and the Trades of the two hemispheres meet without diversion.

In the Western Pacific, the Northern Boundary is still intermittent about the Equator in March. At times during January–March, the Southern Boundary moves to North Australia with the northwesterlies constituting a monsoon which sometimes advances as far as 20° S., undercutting the warm air over the Australian continent so that the boundary slope is upward to the north. The monsoon season is the rainy one in Northern Australia, and Table 9 shows that two-thirds of the annual rainfall at Darwin occurs then.

The northward displacement begins in April, and shows clearly in the mean monthly rainfalls of the islands of Amboina, Palau and Yap (Table 9). By July, the Indian Southwest Monsoon has spread eastward beyond the Philippines and, although the Southern Boundary maintains an east–west orientation along 5° N., the Northern Boundary assumes a north–northwest orientation separating the Indian Monsoon from the North Pacific Trades (Fig. 128*b*).

Because there is mountainous country in larger islands of the Southwest Pacific, the boundaries are frequently distorted and



(a) January

(b) July

Fig. 128 West Pacific: Equatorial Air-stream Boundaries at Low Levels

their seasonal passage not recognisable in the monthly rainfall of places to leeward. Relief also plays a part in developing lines of cloud and rain, the structure of which resembles the 'Sumatras' (page 170). Local cloud-lines of the Pacific form offshore during the

night or early morning and the land breeze is a major factor in their development. Ashton[75] terms them 'coastal fronts', and states that thick altostratus from their tops brings rain to coastal regions during the three hours after dawn. Examples have been reported running east–west on the northern coast of New Guinea, oriented north–south in the Bouganville Strait (Solomon Islands) and across the southern entrance of St. George's Channel in the Bismarck Archipelago.

5. The Central Pacific

Over the Pacific Ocean east of 150° E., the Northeast Trades and the Southeast Trades meet along a single Equatorial Boundary. Intensity of cloud varies greatly from time to time, and rarely produces a continuous line.

The slope of the boundary is probably near-vertical during most of the year, but small temperature-contrasts recorded suggest[76] that the slope is upwards to the north early in the year.

Between 150° E. and 180°, the boundary is oriented east–west, and moves regularly with the seasons, its mean position being 10° N. in August and 10° S. from January to March (Fig. 128*a* and *b*).

East of this region, the equatorial low-pressure trough and the boundary always curve towards the northeast and seasonal movements are slight, the trough never coming farther south than the Equator.[77] Despite this, bad weather is not confined to the northern tropics and there are two semi-permanent lines of disturbed weather in this part of the Pacific: the more northern is without doubt the Equatorial Boundary, because there are consistently northeasterlies on one side of it and southeasterlies on the other. To determine the origin of the other line involves an extra-equatorial genesis, because through a great part of the year it separates two southeasterly streams, each of which flows from the Southern Hemisphere (Fig. 128*b*).

In the Eastern Pacific Ocean, a semi-permanent anticyclone is maintained through the year by a series of migratory anticyclones crossing Australia and the Southwest Pacific. In the trough of low-pressure separating each pair, there is normally an eastward-moving cold front lying along an arc of which a meridian constitutes a chord.[58] To the north of the anticyclonic belt, where the main wind streams are between northeasterly and southeasterly, these cold fronts trail backward to assume an east-southeast orientation. As successive migratory highs go to swell the semi-permanent Eastern Pacific Anticyclone, the trailing prolongations of the meridional fronts maintain a fairly permanent line of disturbed weather

TABLE 9

WEST PACIFIC: MEAN MONTHLY RAINFALL IN INCHES

Place	*Lat.*	*J.*	*F.*	*M.*	*A.*	*M.*	*J.*	*J.*	*A.*	*S.*	*O.*	*N.*	*D.*	*Year*
Guam . . .	13° 29′ N.	4·12	2·45	2·02	1·70	4·21	5·72	12·14	15·24	13·10	10·79	7·68	5·21	84·38
Oloilo . . .	10° 42′ N.	2·35	1·31	0·88	2·28	6·48	10·19	14·09	14·61	13·27	11·66	8·72	5·02	90·86
Yap	9° 29′ N.	5·99	4·76	5·52	7·01	9·90	8·79	16·04	16·06	15·79	14·68	11·19	10·34	126·07
Palau . . .	7° 30′ N.	13·30	5·00	5·30	7·80	13·30	12·30	18·30	15·30	14·00	11·50	12·00	14·00	142·10
Manokwari . .	00° 47′ S.	13·06	7·87	13·27	12·33	7·33	6·81	5·36	5·86	5·75	5·95	6·85	9·17	99·61
Amboina . . .	3° 42′ S.	5·42	4·45	5·20	11·70	22·48	25·44	27·08	22·83	10·44	7·48	3·74	5·24	151·32
Waikaboebak . .	9° 35′ S.	13·70	12·30	14·10	4·80	3·30	0·80	0·70	1·50	1·90	2·60	8·40	13·50	77·60
Darwin . . .	12° 28′ S.	15·87	11·42	12·50	2·29	0·27	0·19	0·00	0·00	0·56	2·11	5·95	7·36	58·52
Apia . . .	13° 48′ S.	23·14	14·46	14·89	9·90	7·15	3·54	3·92	4·42	6·38	8·16	12·34	13·93	122·23
Suva . . .	18° 8′ S.	13·32	11·93	17·09	13·13	12·36	3·78	8·01	7·58	6·48	13·54	12·87	13·96	134·05

TABLE 10

CENTRAL AMERICA: MEAN MONTHLY RAINFALL IN INCHES

Place	*Lat.*	*J.*	*F.*	*M.*	*A.*	*M.*	*J.*	*J.*	*A.*	*S.*	*O.*	*N.*	*D.*	*Year*
San José . . .	9° 56′ N.	0·35	0·20	0·55	2·17	8·70	10·32	8·07	10·64	13·20	13·20	4·84	1·77	74·01
Georgetown . .	6° 50′ N.	9·97	4·49	5·11	6·20	10·79	12·46	10·48	7·85	2·63	3·79	4·15	11·46	89·38
Paramaribo . .	5° 49′ N.	7·90	5·01	7·31	10·08	10·59	11·68	9·45	7·52	3·12	3·36	3·89	7·32	87·23
Quito . . .	0° 13′ S.	4·37	4·86	6·46	6·59	5·56	2·44	0·95	1·14	3·78	5·47	3·85	4·14	49·61
Turi-Assú . . .	1° 43′ S.	5·31	12·79	15·82	15·82	13·41	8·53	6·04	3·17	0·72	0·42	0·63	0·98	83·64
Taperinha . . .	2° 25′ S.	5·33	9·70	14·21	14·19	11·59	7·80	4·21	1·88	1·33	1·90	2·99	3·91	79·04
Manaus . . .	3° 08′ S.	10·92	9·89	12·55	11·26	8·71	3·72	2·05	1·83	2·53	4·39	6·43	8·58	82·86
Senna Madureira .	9° 08′ S.	12·12	9·90	10·32	9·46	4·59	2·67	1·16	2·01	6·54	8·11	8·00	10·32	85·20

south of the Equator and east of 170° E. (YZ of Fig. 128*a* and *b*); this is sometimes called the 'Polar Front'. The Southeast Trades flowing on either side of it are normally of Tropical Maritime origin, and often the more northern stream is turned to a northeasterly early in the year (Fig. 128*a*), when the air to the north might be termed an equatorial stream since it frequently develops conditional instability with increased convection. At such times, the orientations of the streams on both sides of the front indicate convergence, borne out by the variation of mean monthly rainfall at Apia in Samoa (Table 9).

The mean southernmost position of the polar front is slightly north of Suva during March, its incidence showing in the rainfall (Table 9), but the effect is masked because eastern Fiji, being exposed to the Trades, experiences a high frequency of convectional showers throughout the year.

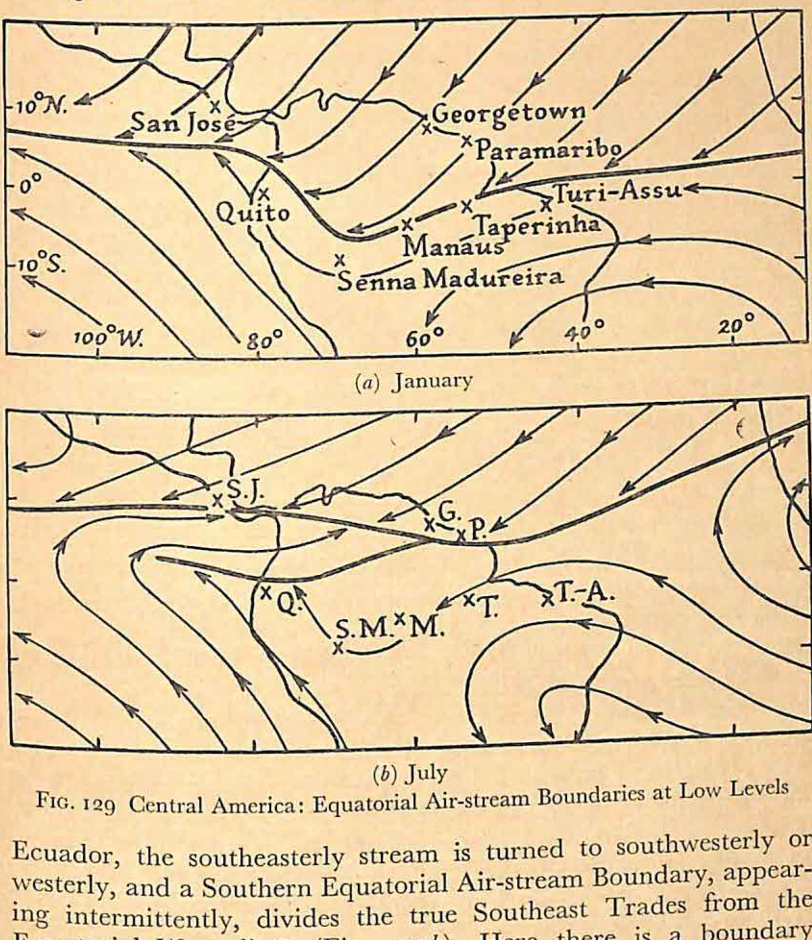
6. South America

During the southern summer, a single air-stream boundary separates the two Trades over Equatorial South America (Fig. 129*a*). Its mean position is about 6° N. on the Pacific coast, and 2° N. on the east and over the Atlantic. The boundary rarely comes more than a few degrees south of the Equator, except inland where it may sometimes reach 10° S. Areas south of the Equator normally have a single indistinct maximum of rainfall January–April (Table 10).

The northward movement of the boundary is rapid after April, and during the latter half of the year its mean position is along the Isthmus of Panama and thence to the east-southeast. It passes northward across the Guiana territories during June and July, and recrosses them when travelling southward during November and December. Evidence of this double passage is seen in the mean monthly rainfall at Paramaribo and Georgetown (Table 10).

The Northeast Trades maintain their northeasterly orientation throughout the year, although during the northern summer they are sometimes temporarily turned in the passing of 'easterly waves'.* The Trades of the Southern Hemisphere, however, undergo considerable diversion in crossing the Southern Continent during the northern summer and may be divided into two separate streams. Air moving from the Atlantic is turned to flow southward over the Brazilian coast, while another stream, probably originating on the eastern side of the Southeast Pacific anticyclone, flows as a southeasterly stream along the coast of Chile. About the latitude of

* For description of 'easterly waves', see Chapter VII.



(*a*) January

(*b*) July

FIG. 129 Central America: Equatorial Air-stream Boundaries at Low Levels

Ecuador, the southeasterly stream is turned to southwesterly or westerly, and a Southern Equatorial Air-stream Boundary, appearing intermittently, divides the true Southeast Trades from the Equatorial Westerlies[66] (Fig. 129*b*). Here there is a boundary structure similar to that of Equatorial Southeast Asia, that of South America being less permanent. During the northern summer it is frequently marked by convection, but during the southern summer it is non-existent, an Equatorial Westerly stream rarely being found.

Coding and Plotting

1. The Code Form for Land Stations

The symbolic form of the international code[5] for synoptic weather stations is as follows:

$iiiT_dT_d$ Nddff VVwwW PPPTT $N_hC_LhC_MC_H$ $(6\underline{D_c}app)$ $(7RR\underline{T_eT_e})$ $(8N_sCh_sh_s)$ $(9S_pS_ps_ps_p)$

Note: The first five groups are always reported.

The groups in brackets are optional, but their presence is identifiable in each case by the first figure which is an indicator.

Those symbols underlined are peculiar to the Southeast Asian region. In other parts of the world, different elements occupy these positions.

A specimen coded report is on p. 210.

2. The Elements of the Code

The meanings of the separate symbols are as follows:

iii — Station Number: Each station has a separate number from which the origin of a report may be determined. See Table I for the numbers of selected stations in Equatorial Southeast Asia.

T_dT_d — Dew-point Temperature in whole degrees Fahrenheit (some countries report in degrees Centigrade).

N — Total Amount of Cloud in eighths of the sky covered. i.e. N = 0 is clear sky, N = 8 is overcast and N = 9 signifies that the sky is obscured or that the cloud amount cannot be estimated because of darkness.

dd — True Direction (in tens of degrees) from which the wind is blowing. dd = 00 is calm; dd = 01 signifies that the wind is from 010° true; dd = 18 denotes a wind from 180° (i.e. a southerly).

ff — Wind speed in knots.

VV	Visibility according to the scale in Table II; here only certain values have been included from the complete visibility scale.
ww	Present Weather according to Table III.
W	Past Weather according to the scale of Table IV.
PPP	Pressure in millibars reduced to mean sea-level; including the tenths of millibars but omitting the hundreds or thousands. Thus PPP = 103 represents a pressure of 1010·3 millibars; PPP = 964 is a pressure of 996·4 millibars.
TT	Air Temperature in whole degrees Fahrenheit (or Centigrade).
N_h	Amount of the lowest cloud, the height of which is reported by h. N_h is in the same scale as N.
C_L	Type of low cloud according to Table V.
h	Height above ground of the base of the lowest cloud according to Table VI.
C_M	Type of middle cloud according to Table VII.
C_H	Type of high cloud according to Table VIII.
6	The indicator showing that this optional group is being reported.
D_c	Direction from which the lowest cloud is moving according to a scale of eight points; i.e. $D_c = 1$ means from northeast; $D_c = 2$ is from east; $D_c = 8$ is from true north; $D_c = 0$ signifies that the cloud is stationary, and $D_c = 9$ denotes that the movement is variable.
a	Characteristic of the Barometric Tendency in the three hours preceding the time of observation. The way in which the barometer has risen or fallen is described by this figure according to Table IX.
pp	Barometric Tendency: The amount, in tenths of millibars, that the barometer has risen or fallen during the last three hours; i.e. pp = 14 represents a change of 1·4 millibars, and the sense of the change is shown by the code figure for a.

7 An indicator showing that this optional group is being reported.

RR Amount of precipitation according to Table X. Rainfall is normally measured in inches, and the reading must be converted to millimetres by multiplying by 25·4. It will be noted from Table X that, in the most probable range of rainfall, the code figure is identical with the actual rainfall in millimetres.

The procedure in Southeast Asia is that
(*a*) at 0000 hours (G.M.T.), RR indicates the amount of rainfall in the preceding twenty-four hours, but
(*b*) at all other synoptic reporting hours, RR represents the amount of rainfall since 0000 hours (G.M.T.).

T_eT_e The Maximum or Minimum Temperature of the day in whole degrees Fahrenheit (some countries report in degrees Centigrade). The reporting of this code element is peculiar to the region of Southeast Asia.

8 An indicator showing that this optional group is being reported.

N_s Amount of the Significant Cloud Layer in eighths. The significant cloud-layers are:
(1) The lowest layer of cloud below 20,000 ft. covering more than half the sky.
(2) The lowest layer of cloud, if any, below the layer given in the preceding specifications.
(3) If no layer of cloud below 20,000 ft. covers more than half the sky, the significant cloud layer is the lowest layer of cloud below 20,000 ft.

C Types of Significant Cloud according to Table XI.

h_sh_s Height above Station of the Base of the Significant Cloud according to Table XII. If the cloud base is between two of the heights given in the Table, the code figure for the lower height is reported.

$9S_pS_ps_ps_p$ This group is used to report special phenomena such as directions of clouds and particulars of fog banks and of thunderstorms. (Details may be found in Fascicule 1, Meteorological Codes, Publication No. 9, Organisation Météorologique Internationale.)

3. Code Tables

Table I: Station Numbers—iii

Thailand	564 Phuket
Singapore and Malaya	601 Penang
	603 Alor Star
	615 Kota Bharu
	619 Kuala Trengganu
	620 Sitiawan
	647 Kuala Lumpur
	657 Kuantan
	660 Port Swettenham
	665 Malacca
	674 Mersing
	694 Singapore (Kallang)
Sarawak and North Borneo	413 Kuching
	465 Labuan
	471 Jesselton
	491 Sandakan
Sumatra	001 Sabang
	035 Medan
	163 Padang
	221 Palembang
	241 Banka
	249 Billiton
Borneo	509 Tarakan
	583 Pontianak
	633 Balikpapan

Table II: Visibility—VV

Code Figure	*Distance*	*Code Figure*	*Distance*
X1	20 yds.	16	2 mls.
X3	60 ,,	24	3 ,,
X5	100 ,,	32	4 ,,
00	$<$220 ,,	40	5 ,,
01	220 ,,	48	6 ,,
02	440 ,,	80	10 ,,
04	880 ,,	81	12 ,,
08	1 ml.	82	24 ,,
10	$1\frac{1}{4}$ mls.	83	36 ,,

Table III: Present Weather—ww

ww = 01 to 19: No precipitation, fog, duststorm, sandstorm or drifting snow at the station at the time of observation or during the preceding hour.

Code Figure	*Meaning*
00	Cloud development not observed.
01	Clouds generally dissolving or becoming less developed.

02	State of sky on whole unchanging.
03	Clouds generally forming or developing.
04 to 09	Details of haze, smoke or sand reducing visibility.
10	Mist.
11 to 12	Shallow fog.
13	Lightning visible but no thunder heard.
14 to 16	Precipitation within sight but not at station.
17	Thunder heard but no precipitation at station.
18	Squalls during last hour.
19	Funnel clouds (water-spouts, etc.) within sight during last hour.

ww = 20 to 29: Phenomena at station during preceding hour but not at time of observation.

Code Figure	*Meaning*
20	Drizzle.
21	Rain.
22 to 24	Snow or freezing rain.
25	Showers.
26 to 27	Snow and hail showers.
28	Fog.
29	Thunderstorm, with or without precipitation.

ww = 30 to 39: Duststorm, sandstorm or drifting snow.

Code Figure	*Meaning*
30	Slight or moderate duststorm or sandstorm, decreasing.
31	,, ,, ,, ,, ,, ,, unchanging.
32	,, ,, ,, ,, ,, ,, increasing.
33	Severe duststorm or sandstorm, decreasing.
34	,, ,, ,, ,, unchanging.
35	,, ,, ,, ,, increasing.
36 to 39	Drifting snow.

ww = 40 to 49: Fog at time of observation.

Code Figure	*Meaning*
40	Fog in distance.
41	,, ,, patches.

42	Fog becoming thinner, sky discernible.
43	,, ,, ,, ,, not discernible.
44	,, unchanging, sky discernible.
45	,, ,, ,, not discernible.
46	,, becoming thicker, sky discernible.
47	,, ,, ,, ,, not discernible.
48 to 49	,, depositing rime.

ww = 50 to 59: Drizzle at station at time of observation.

Code Figure	*Meaning*
50	Drizzle, slight intermittent.
51	,, ,, continuous.
52	,, moderate, intermittent.
53	,, ,, continuous.
54	,, thick, intermittent.
55	,, ,, continuous.
56 to 57	,, freezing.
58	,, and rain, slight.
59	,, ,, ,, moderate or heavy.

ww = 60 to 69: Rain at station at time of observation.

Code Figure	*Meaning*
60	Rain, slight, intermittent.
61	,, ,, continuous.
62	,, moderate, intermittent
63	,, ,, continuous.
64	,, heavy, intermittent.
65	,, ,, continuous.
66 to 69	,, freezing or with snow.

ww = 70 to 79: Solid precipitation (such as snow) not in showers.

ww = 80 to 99: Showery precipitation, or precipitation with recent or current thunderstorm.

Code Figure	*Meaning*
80	Rain showers, slight.
81	,, ,, moderate or heavy.

82	Rain showers, violent.
83 to 90	Showers of snow or hail with or without rain.
91	Thunderstorm during preceding hour; slight rain at time of observation.
92	Thunderstorm during preceding hour; moderate or heavy rain at time of observation.
93 to 94	Thunderstorm during preceding hour; snow or hail at time of observation.
95	Thunderstorm, slight or moderate, with rain at time of observation.
96	Thunderstorm, slight or moderate, with hail at time of observation.
97	Thunderstorm, heavy, with rain at time of observation.
98	Thunderstorm, with duststorm or sandstorm at time of observation.
99	Thunderstorm, heavy, with hail at time of observation.

Note: In general the highest applicable code figure is used.

Table IV: Past Weather—W

Code Figure	*Description*
0	Clear, or few clouds.
1	Variable sky.
2	Cloudy or overcast.
3	Sandstorm, duststorm or drifting snow.
4	Fog or thick dust haze.
5	Drizzle.
6	Rain.
7	Snow.
8	Shower(s).
9	Thunderstorm.

Table V: Types of Low Cloud—C_L

Code Figure	*Description*
0	No low cloud.
1	Fair weather cumulus (cumulis humilis)
2	Cumulus of considerable development, generally towering.
3	Cumulonimbus with indistinct tops which are not cirriform or anvil-shaped.

4	Stratocumulus formed by the spreading out of cumulus.
5	Stratocumulus not formed by the spreading out of cumulus.
6	Stratus and/or fractostratus, but not fractostratus of bad weather.
7	Fractostratus and/or fractocumulus of bad weather (scud), usually under altostratus and nimbostratus.
8	Cumulus and stratocumulus other than those formed by the spreading out of cumulus; with bases at different levels.
9	Cumulonimbus having a cirriform top, often anvil-shaped.

Table VI: Height of Lowest Cloud—h

Code Figure	*Height in Feet*
0	0– 150
1	150– 300
2	300– 600
3	600–1000
4	1000–2000
5	2000–3500
6	3500–5000
7	5000–6500
8	6500–8500
9	No low cloud.

Table VII: Types of Middle Cloud—C_M

Code Figure	*Description*
0	No middle cloud.
1	Thin altostratus.
2	Thick altostratus or nimbostratus.
3	Thin altocumulus, cloud elements not changing much; at a single level.
4	Thin altocumulus in patches; cloud elements continually changing and/or at more than one level.
5	Thin altocumulus in bands or in a layer gradually spreading over the sky and thickening.
6	Altocumulus formed by the spreading out of cumulus.
7	Altocumulus and altostratus at same or at different levels.
8	Altocumulus castellatus.
9	Altocumulus of a chaotic sky, generally at different levels; dense cirrus patches are usually also present.

Table VIII: Types of High Cloud—C_H

Code Figure	*Description*
0	No high cloud.
1	Strands of cirrus, scattered and not increasing (often 'mares' tails').
2	Dense cirrus in patches or twisted sheaves usually not increasing.
3	Cirrus, often anvil-shaped, either the remains of a cumulonimbus or portions of an existing distant cumulonimbus.
4	Cirrus (often hook-shaped) gradually spreading and thickening.
5	Cirrostratus, with or without cirrus; gradually spreading and thickening, but the continuous layer not reaching 45° altitude.
6	Cirrostratus, with or without cirrus; gradually spreading and thickening, and the continuous layer exceeding 45° altitude.
7	Cirrostratus covering the whole sky.
8	Cirrostratus (with or without cirrus and cirrocumulus), not increasing and not covering the whole sky.
9	Mainly cirrocumulus (with or without some cirrus or cirrostratus).

Table IX: Characteristic of Barometric Tendency—a

Code Figure	*Specification*
0	Rising, then falling.
1	Rising, then steady; or rising, then rising more slowly.
2	Unsteady.
3	Steady or rising.
4	Falling or steady, then rising; or rising, then rising more rapidly.
5	Falling, then rising.
6	Falling, then steady; or falling, then falling more slowly.
7	Unsteady.
8	Falling.
9	Steady or rising, then falling; or falling, then falling more rapidly.

Note: In code figures 0–4, the present atmospheric pressure is higher or the same as three hours previously. In code figures 5–9, the present atmospheric pressure is lower than three hours previously.

Table X: Amount of Rainfall—RR

Code Figure	Specification	Code Figure	Specification
00	Nil	91	0·1 millimetre
01	1 millimetre	92	0·2 "
02	2 millimetres	..	....
..	....	..	....
..	....	95	0·5 millimetre
54	54 millimetres	96	0·6 "
55	55 "	97	Trace of rain
56	60 "	99	Unreliable measurement
57	70 "		
..	....		
60	100 millimetres		

Table XI: Types of Significant Cloud—C

Code Figure	Type	Code Figure	Type
1	Cirrus	6	Stratocumulus
2	Cirrostratus	7	Nimbostratus
3	Cirrocumulus	8	Cumulus or fractocumulus
4	Altocumulus	9	Cumulonimbus
5	Altostratus	0	Stratus or fractostratus

Table XII: Height of Base of Significant Cloud—h_sh_s

Code Figure	Height in Feet	Code Figure	Height in Feet
00	<100	81	9000
01	100	83	10,000
02	200	84	13,000
03	300	85	16,000
..	...	86	20,000
..	...		
79	7900		
80	8000		

4. Specimen Coded Reports

The following are two examples of coded reports and the decoded plain-language versions. For simplicity all the optional groups have been omitted.

	$iiiT_dT_d$	Nddff	VVwwW	PPPTT	$N_hC_LhC_MC_H$
(*a*)	61576	80415	32638	09786	6732x
(*b*)	00172	61802	82030	11280	42506

The meaning of report (*a*) is as follows:

Station: Kota Bharu

Dew-point: 76° F.

Total Cloud Amount: 8 eighths

Wind Direction: from 40°

Wind Speed: 15 knots

Visibility: 4 miles

Present Weather: Moderate continuous rain

Past Weather: Showers

Pressure: 1009·7 millibars

Temperature: 86° F.

Amount of Low Cloud: 6 eighths

Type of Low Cloud: Fs or Fc of bad weather

Height of Low Cloud: 600 to 1000 ft.

Type of Middle Cloud: Thick As or Ns

Type of High Cloud: Unknown

The meaning of report (*b*) is as follows:

Station: Sabang

Dew-point: 72° F.

Total Cloud Amount: 6 eighths

Wind Direction: from 180°

Wind Speed: 2 knots

Visibility: 24 miles

Present Weather: Clouds generally forming or developing

Past Weather: Clear or few clouds

Pressure: 1011·2 millibars.

Temperature: 80° F.

Low Cloud Amount: 4 eighths

Low Cloud Type: Towering Cu

Height of Low Cloud: 2000 to 3500 feet

Middle Cloud Type: Nil

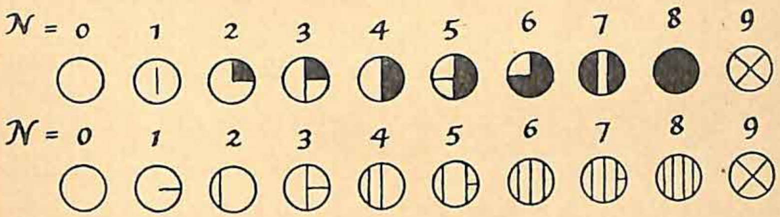
High Cloud Type: Cs spreading and thickening and exceeding 45° altitude

5. Plotting

The procedure for plotting the synoptic chart is as follows:

(*a*) Find the position of the reporting station on the blank synoptic chart and draw a circle of approximate diameter $\frac{1}{8}$ to $\frac{3}{16}$ inch.

(*b*) Inside the circle mark in the total cloud amount (N) according to either of the scales shown in Fig. i. (In the examples herein the second scale is adopted.)



i Symbolisation of N

(*c*) Draw an arrow towards the station circle from the direction of the wind.

(*d*) On the end of the arrow make ticks to denote the Beaufort Wind-force, a small tick for Force 1 and a longer tick for Force 2. The coded version gives wind speed in knots so that, before plotting, this must be converted to Beaufort Force as follows:

Speed, knots . .	2	5	9	14	19	24	30	37	44	52	60
Beaufort Force . .	1	2	3	4	5	6	7	8	9	10	11

(*e*) The remaining elements of the report are then grouped around the circle in the order shown in Fig. ii. Symbols are used for plotting ww, W, C_L, C_M, C_H (see Fig. iv); the other elements are plotted in the code figures.

C_H

TT C_M PPP

VVww(N)

T_dT_d $C_L N_h$ W

h

ii Plotting Model

(*f*) Present-weather symbols corresponding to the various code numbers (ww = 00 to 99) are shown in Fig. iv. These are always plotted just to the left of the circle, except when ww = 00, 01, 02 or 03. In the latter cases, no symbol is inserted in the normal position for ww, but certain strokes are made about the station circle as shown in Fig. iv.

(*g*) It is advantageous to plot certain elements in black and others in red, but in the examples shown herein, one colour has been used throughout.

6. Examples of Plotted Reports

The plotted versions of the coded reports previously given for Kota Bharu and Sabang are shown in Fig. iii.

iii Plotted Weather-reports

(*a*) Kota Bharu (*b*) Sabang

7. Projected Code Changes

At the 1953 Session of the Commission for Synoptic Meteorology of World Meteorological Organisation, the following code form was recommended[78] for general adoption:

iii Nddff VVwwW PPPTT $N_hC_LhC_MC_H$ $T_dT_dj_aj_pj_p$ (7RRjj)
($8N_sCh_sh_s$) ($9S_pS_ps_ps_p$)

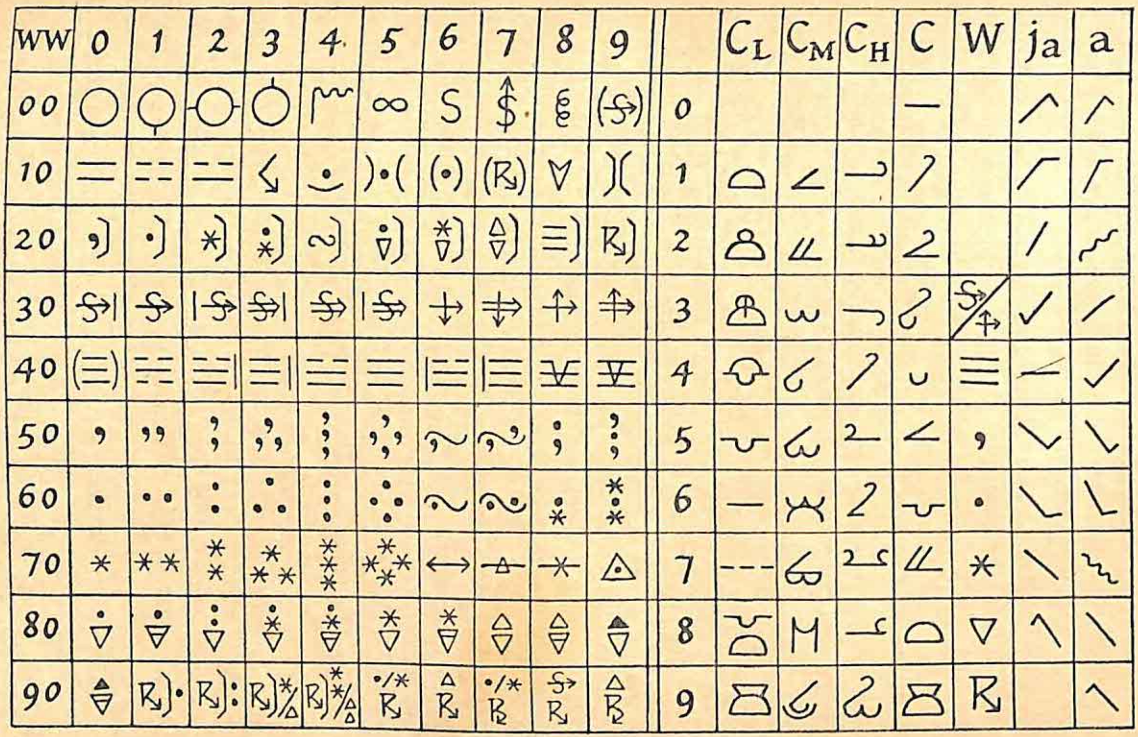
Note: $j_aj_pj_p$—When $j_a = 0 - 8$, this designates the barometric tendency according to Table XIII. Then j_pj_p gives the value of the change in the last three hours in tenths of millibars. When $j_a = 9$, the meaning of j_pj_p is as fixed by international agreement.

jj—Representation fixed by international agreement in different regions of the world.

VV—Code in Table XIV.

h_sh_s— ,, ,, ,, ,, .

plotted reports, a small tick on the wind-arrow should denote 5 knots and a large tick 10 knots. It was also decided that, in



iv Symbols for Plotting Individual Observations

Table XIII: Characteristic of Barometer Tendency—ja

Code Figure	Specification
0	Increasing, then decreasing.
1	Increasing, then steady; or increasing, then increasing more slowly.
2	Increasing (steadily or unsteadily).
3	Decreasing or steady, then increasing; or increasing, then increasing more rapidly.
4	Steady or unsteady, atmospheric pressure the same as three hours ago.
5	Decreasing, then increasing.
6	Decreasing, then steady; or decreasing, then decreasing more slowly.
7	Decreasing (steadily or unsteadily).
8	Steady or increasing, then decreasing; or decreasing, then decreasing more rapidly.

Note: In code figures 0–3, the present atmospheric pressure is higher than three hours previously.

In code figures 5–8, the present pressure is lower than three hours previously.

Table XIV: Visibility (VV) and Height of Significant Cloud (h_sh_s)

Code Fig.	VV in yds.	h_sh_s in ft.	Code Fig.	VV in mls.	h_sh_s in ft.	Code Fig.	VV in mls.	h_sh_s in ft.
00	<110	<100	51 to 55 not used			82	25	40,000
01	110	100	56	$3\frac{3}{4}$	6000	83	$28\frac{1}{8}$	45,000
02	220	200	57	$4\frac{3}{8}$	7000	..	..	.....
03	330	300	..	...	....	89	$>43\frac{3}{4}$	$>70,000$
04	440	400	78	$17\frac{1}{2}$	28,000			
05	550	500	79	$18\frac{1}{8}$	29,000			
..	..	...	80	$18\frac{3}{4}$	30,000			
50	5500	5000	81	$21\frac{7}{8}$	35,000			

APPENDIX B

Symbols on Synoptic Charts

A. Air-stream Boundaries:

(1) Convergence Line. A particular case of air-stream boundary when convergence is assumed from cumuliform build-up.

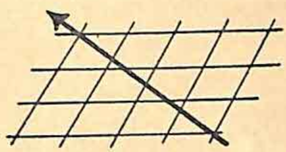
(2) Air-stream Boundary with stratiform cloud.

(3) Inactive Air-stream Boundary or one on which types of cloud unknown.

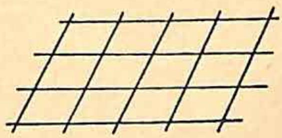
B. Disturbances within a Stream:

(1) Convergence Zones

—Moving

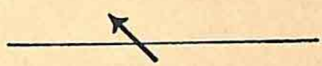


—Stationary



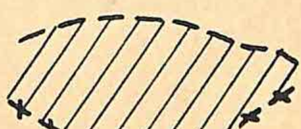
(2) Shear Lines

—Moving

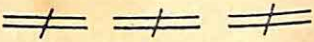


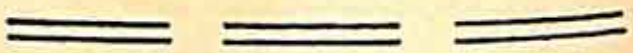
—Stationary

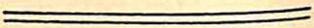
C. Doldrums: Doldrum edges are marked by the appropriate line-symbol from Scale A above.



D. Intertropical Front: Outside latitudes 10° North and 10° South a double line is drawn, cloud structure being denoted by the method employed with Air-stream Boundaries:

(1) Cumuliform 

(2) Stratiform 

(3) Inactive 

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Index

EQUATORIAL WEATHER

THIS book has evolved from the author's long experience as an equatorial meteorologist and lecturer at the University of Malaya. Its presentation of the principles of equatorial weather analysis will make it a valuable work of reference for the student of meteorology and for those concerned with transport over tropical oceans and air routes.

The keynote to the general scheme of analysis is continuity and, although there are many references to Southeast Asia, it is applicable to the whole of the equatorial belt. Those techniques which are of proven worth are competently described with the help of detailed examples. The work includes chapters on climate, prevailing wind-streams at all levels and the diurnal variation and distribution of cloud and rain.

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